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Marine Hydrometeorological Information and Forecasting

by N. A. Belinskiy

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PREFACE

In the past few years a considerable amount of research had been conducted in the field of the development of the methods of marine hydrometeorological forecasting, which touch upon a large number of questions. While in the beginning of the development of the work of forecasting in the marine field a large portion of research was devoted to the development of methods for marine ice forecasting, and for the most part long-range forecasting, in recent times the development of the methods has been carried on on a somewhat larger scale, touching upon a large number of elements of the marine system. Research is being conducted not only in the development of long-range forecasting methods, but also in the field of short-term forecasting. In the development of short-term forecasting methods, we have the opportunity of watching step by step the development of the processes which occur at sea in the concrete physiogeographical system, widely utilizing known physical principles. Unfortunately, we cannot always succeed in this when working on methods of long-term forecasting in the marine field. In such a case we usually build hypotheses which are then proved by purely statistical methods. An insufficient amount of ~~properly placed research~~ frequently ~~does not~~ allow us to make a strictly statistical check of the propounded hypotheses, with the result that the question of the quality of the long-range forecasting method is sometimes decided only by the material of current research. Not infrequently, the originally received research results prove unsatisfactory after a few years which hinders the continuous and large work of developing the method.

We should regard the development of the method of marine hydrometeorological forecasting as a way of applying general laws of the physics of the sea and the atmosphere to the concrete physiogeographical system. This circumstance requires parallel research on short-term and long-range forecasting in the marine field. With such a situation the researcher can make extensive use of the method of physiomathematical analysis as a basic method of research, and as an auxiliary method, for tabulating and evaluating some of the relations obtained -- the statistical method.

In this book are presented some results of research in the field of marine hydrometeorological forecasting made known largely after the First All-Union Conference of Specialists in Marine Forecasting, which met in 1938 at the State Hydrological Institute. Its aim was -- not so much to supply research data in this field, as to give an objective account of the status and possibilities of the methods and most acute tasks ahead in the light of developments, which appear to be most in prospect for the near future.

The book chiefly examines methods, that are basic in the analysis of physical processes, extended to the sea, applicable in any geographical situation, understandable when taking account of local peculiarities. The very subject, to which the book is dedicated, is such, that it obligates us to refer constantly to data derived from observation. Therefore, there is in the book a clearly expressed leaning in the direction of empirical studies. At the same time the author does not entirely consider that the only possible road in this field is the road of empirical studies. The indicated inclination is explained in that the overwhelming majority

of the work in this field carried just this empirical character. This course of the work, probably, should be recognized as being legitimate in view of the limited time since the beginning of the development of research in this field, and also because of the applied meaning of the forecasts.

The limited scope of the book has forced the author to give in the first chapter in an extremely condensed form a general presentation of the roads taken by the development of the methods of marine forecasting, merely noting the knotty problems. It has far from fully listed the names of researchers working in this field. This gap is filled by the contents of the book. Everywhere in the pertinent sections the author indicates by whom the research was carried out in a particular field. It is characteristic that most of the names mentioned in the text are those of young researchers, which is natural, since the work in the field of marine forecasting was begun just recently. During the past few years the centers for the research and operational work in marine forecasting have been the Central Institute of Forecasting and the Arctic Scientific Research Institute.

In the well known work of V. Yu. Vize "The Basic Methods of Long-Range Ice Forecasting in Arctic Waters", honored by a Stalin Prize, a great many studies in the field of long-range ice forecasting which were carried on mainly in the Arctic Scientific Research Institute are summarized. Therefore, it was inexpedient to refer to the work here which was studied in it, bearing in mind the limited scope of the book. This is the reason why the book shows preference to the work of the collaborators of the Central Institute

of Forecasting. To a certain extent the fact that the author himself is a collaborator of the Central Institute of Forecasting and ^t took part in much of the work as a director or an executive, could not fail to influence the selection of the material.

The book itself is written as a training aid available to as large a group of students as possible and as an aid to the practical workers in the field of marine forecasting. This could not fail to leave its mark. The book is far from utilizing all the literature. Here only the theoretical studies are mentioned which help us to understand the substance of the observed phenomena or which may be used in forecasting. Only those empirical studies were used which can be useful in the practical forecasting or are interesting with respect to methodology.

Work on this book presented several difficulties, connected in the first place with the fact that there were no similar publications in Soviet or foreign literature. Therefore, naturally the book is not without shortcomings. Nevertheless, the publication of the book is in keeping with the necessity of developing and further expanding research in the field of marine hydrometeorological forecasting methods. The development of this work is necessary due to all the growing demands of the national economy for quality, contents and scope in forecasts of the state of the sea.

CHAPTER I

THE DEVELOPMENT OF THE METHODS OF MARINE
HYDROMETEOROLOGICAL FORECASTING

1. SOME INFORMATION ON THE RISE AND DEVELOPMENT OF
 HYDROMETEOROLOGICAL FORECASTING

Marine hydrometeorological forecasting based on scientifically developed methods has a comparatively short history. Only in the beginning of the 20th century did articles appear in which the first attempts at developing methods of ice forecasting were set forth. At the same time, man using the sea for ^mavigation, fishing, and whaling could not forecast for himself nor his comrades to aid in the industry and travel. The sea is so changeable that even today despite the tremendous experience gained by mankind and the vigorous development of technology, it is not uncommon for people to pay with their lives for the poor knowledge of the sea or for ^megligence. Setting out to sea a mariner cannot ignore what might happen to him enroute, and in thinking of this he builds an assumption based on conditions at sea that he is observing and the experience and knowledge gained by him previously. The more experience and knowledge we have about the sea, naturally the more substantiated appear these suppositions and the more often these suppositions correspond to the realities. That is why among the most backward, and among civilized peoples alike experienced mariners who know the caprices of the elements and know how to fight them are highly valued and distinguished. At the same time even an experienced marine^r, well acquainted with the sea, cannot foretell all its caprices.

In seafaring history many brilliant forecasts are known, which were made without any specially worked out methods but were based on careful analysis of physiogeographical data. For such a forecast we must refer to the brilliant forecast of the great Russian scholar M. V. Lomonosov, who foresaw the possibility of navigating the northern sea route around Eurasia. The world is much indebted to the Russian and Soviet mariners and scholars for their discoveries about the Arctic and Antarctic. Russian navigators led by Bellinsgauzen and Lazarev were the first to open the Antarctic continent. The Soviet people brought to reality the navigation around Eurasia as predicted by Lomonosov and made this route a regularly travelled seaway. The icebreaker invented by the Russian scholar-navigator Admiral S. O. Makarov served this undertaking. The battle against the harsh elements of the Arctic gave top priority to problems of developing ice forecasting. The first ^{direct} ~~attempt~~ attempt at developing a special method of forecasting the position of the southern edge of the ice of the Barents Sea was made in 1913 by E. Lesgaft. He published in Zapiskakh po Gidrografii (Hydrographical Notes) an article "Ices of the Kara Sea and Its Accessibility for Communications with Siberia", in which he brings out the relationship between the winter temperatures of water and air and the position of the southern edge of ice in the Barents Sea from May to June.

In addition V. Yu. Vize did a great deal of work on the development of marine ice forecasting methods. Beginning in 1923 he issued forecasts of the ice conditions in the Barents Sea, and, later on in the Kara Sea.

Lesgaft believed that the ice conditions in the Barents Sea were determined by the temperatures of the water and air. Vize,

as a result of his observations, came to the conclusion that "in the formation of ice conditions in any area special importance must be ascribed to the meteorological peculiarities, particularly the air pressure in the regions located at a great distance around the body of water whose ice conditions are being studied." Developing this idea further, Vize contended that the variations in the ice conditions in individual arctic seas occur as a result of the variations which occur in the general circulation of the atmosphere. For the characterization of the peculiarities of the atmospheric circulation of various years Vize used two procedures: establishment of barometric scales, (Barometric scales are charts of mean pressures of the atmosphere. To compile them, all the years, for which a considerable deviation from the normal ice conditions in the Barents Sea was observed, were divided into two groups; those with a heavy and a light ice condition. The atmospheric pressure was averaged separately for a few months preceding heavy and light ice conditions. Furthermore, developing the methods of barometric scales of V. Yu Vize, G. Ya. Vangengeim arrived at the typifications of the atmospheric processes and perfected the method of forecasting the ice condition in the Barents Sea.) and of correlations of the quantitative characteristics of ice phenomena with atmospheric pressures at several selected points which were often dispersed over the entire earth.

Ice forecasting acquired especially great importance after the Main Administration of the Northern Sea Route was organized, which had been given the task of converting the Northern Sea Route into a regularly travelled seaway. In connection with this, an interdepartmental bureau of ice forecasting was created with the mission

of developing methods of marine ice forecasting in arctic waters. This bureau attracted the best specialist^S in navigation. The attraction of a large circle of specialists brought about the development and broadening of the basic principles, on which the methods of marine ice forecasting were to be based.

V. Yu. Vize, contending that variations in the ice conditions in arctic waters depended on atmospheric processes, denied the possibility of a direct influence of the warm Atlantic waters emptying into the North Arctic Ocean upon the ice conditions of the sea lying to the east of the Barents Sea. V. V. Shuleykin and N. N. Zubov considered this point of view erroneous and attached great importance to the warm Atlantic waters flowing into the arctic seas. This question was correctly answered only after the expedition of the hydrographical ship Traymyr into the Kara Sea. This expedition was participated in by V. V. Shuleykin who set up a series of observations permitting the subsequent calculation of a complete heat balance for the Kara Sea. The calculations showed that the warm Atlantic waters played a tremendous role in the life of the sea. The Kara Sea would freeze through completely if the Warm Atlantic waters did not flow into it. Further expeditionary studies fully confirmed the correctness of this conclusion made on the basis of the computations. G. E. Ratmanov succeeded in finding warm Atlantic waters even in the northern part of the Chukchi Sea.

The work of V. V. Shuleykin in the calculation of the thermal balance of the Kara Sea was ^a turning point in the development of the methods of marine ice forecasting. After this work it became apparent that a genuinely scientific method of ice forecasting must be built on the basis of a careful study of the

geneses of the predicted phenomenon, that is, the sufficiently substantiated forecasts of ice conditions can be made only if the possibility of predicting changes in the basic elements of the heat balance presents itself.

During the years of the First Five-Year Plan the training of cadres of hydrometeorologists went far. In 1930 a special Hydrometeorological Institute was created whose task it was to train Hydrometeorological specialists who would be well equipped with physio-mathematical and geographical knowledge. In this same period a new branch of knowledge was created -- the physics of the sea. In 1933 V. V. Shuleykin published a book Fizika morya [Physics of the Sea] in which all the most important physical phenomena were examined which are observed in the sea or are connected with the life of the sea. The training of specialists capable of learning the physical processes at sea, and the appearance of literature responding to this trend could not help but leave a mark in the development of more intensified research in the field of developing methods of marine hydrometeorological forecasting.

In 1937 within the Central Weather Bureau in Moscow a small section was organized, which was entrusted with carrying on the operational work of marine forecasting to serve the demands of the central economic and defense organizations (one of the first collaborators of this section was M. M. Somov). To guarantee marine forecasts for local organizations rested with the specially created groups of marine forecasting within the local Weather Bureaus. In this period A. I. Karakash began compiling short-term forecasts of the tides and currents for the Caspian Sea thereby reviving the work begun by N. N.

Struyskiy in his time.

Scientific research in the field of marine forecasting was begun about the same time in the marine sections of the Government Hydrological Institute too. In 1938 it brought together the First All-Union Conference of Marine Hydrometeorological Forecasting.

After this Conference the development of the methods of forecasting broadened: variations in the tides, disturbances, flow and temperatures of the water. The methods of ice forecasting for non-arctic waters actually began progressing also from this time on.

Marine forecasting is progressing considerably slower abroad than it is here in the USSR. We can mention only a few foreign efforts which have some bearing on the development of the methods of marine ice forecasting. There are no definite foreign methods of ice forecasting even today, not counting the purely empirical method of forecasting icebergs in the Newfoundland Banks region. There are also some empirical connections for forecasting non-periodic variations in the tides.

CHAPTER II

THE HYDROMETEOROLOGICAL SERVICES FOR THE NATIONAL
ECONOMY

Hydrometeorological services to the national economy consist in the provision of information on the long-term hydrometeorological regime, current data on the condition of the sea and weather, and of forecasting weather and the condition of the sea for the period ahead.

2. INFORMATION ABOUT THE HYDROMETEOROLOGICAL REGIME

To answer questions about the hydrometeorological regime and the climate of the seas, the results of observation by a network of stations and observatories, of expeditions and routine observations made by ships and aircraft are published. This news is published in the form of statistics in yearly publications or other similar publications. In these publications we can find not only the product of direct observations, but also some conclusions therefrom: mean values, extreme values, values of frequencies of one phenomenon or another. Besides, the local hydrometeorological administrations, observatories, the State Oceanographic Institute, and the Main Geophysical Observatory own the archives of hydrometeorological observations, from which we can draw up data about the results of observations made by a former network of marine stations, observatories or expeditions. Recently the Central Meteorological Archives were organized in Moscow, in which hydrometeorological observations, whenever they may have been made, are gradually being concentrated.

Useful information about the regime can also be derived from the efforts of scientific research institutes (FOIN, GGO and others), from articles in various magazines (Meteorologiya i gidrologiya, Zapiski po gidrografii, etc)

As a result of processing the products of observations we may obtain the answers to the questions: what is the average long-range weight of this or that element of the regime; what is its extreme weight. Maybe it may also be established, what the frequency or probability of individual phenomena or conditions in the regime. Finally, from the products of direct observations we may solve the question: what hydrometeorological conditions were being observed in a given area at a certain time.

To obtain the answer to this or other questions with the aid of the products of observations, the specialist must spend considerable time on processing statistics. In many cases this is impractical and sometimes even impossible, as, for example, on board ships, in wartime, etc. In connection with this, except for the statistics obtained through observations, hydrological manuals are published about seas and separate areas of seas, the hydrometeorological characteristics of the seas and oceans, hydrometeorological outlines for pilots, hydrometeorological maps of the sea^s, physiogeographical atlases, etc. The more important information is placed in these publications, mostly that which is necessary for navigators in commercial and military organizations.

It is not necessary to dwell on the contents of these publications in more detail here. We must always remember, that any hydrometeorologist who must provide the national economy with forecasts, facts or information about of this or that sea or on a

separate area of a sea, must carefully acquaint himself with all the reference works and above all with the above-mentioned literature. The quick, and also the high quality, attention to the demands of the national economy with information about phenomena of the sea can be made possible only ^{by d} ~~goos~~ familiarity with the regime of the sea, and consequently, with the reference literature on the subject.

Aside from reference literature, it is most desirable to use data of direct observations. This data is necessary for two reasons. First, reference literature cannot as a rule provide the complete answers to all the demands of the national economy. Secondly, only with the aid of data from observations made in recent years can we verify a certain method of forecasting or set up the new numerical and forecasting relations required to serve the national economy with forecasts.

3. INFORMATION ABOUT THE CURRENT SEA CONDITION

Data on the current hydrometeorological conditions are required more often for the national economy and in larger volume than information on the marine hydrometeorological regime. To provide this information is a very complicated and extensive task.

..... Most of the hydrometeorological stations are engaged in the job of obtaining data. In addition, the shipboard hydrometeorological stations organized on ships of the Ministry of Merchant Fleet and on ships of the Ministry of Fishing Industry are also devoted to the job of obtaining data. When expeditionary vessels are at sea or when air reconnaissance is being carried on, the observations of the expeditionary vessels and aircraft are also used widely for a complete

illumination of the condition of the sea at a particular point in time. Shore, island, shipboard (routine) and expeditionary stations make hydrometeorological observations not less than four times per day. These observations are made in accordance with special instructions and regulations. Immediately after making the observations the results are transmitted by telegraph or by radio at strictly fixed times to the hydrometeorological administrations in accordance with the special radiotelegraph or telegraph schedules. In order to give the most complete information possible to the hydrometeorological administrations in a minimum number of words in the telegrams, codes especially developed for this purpose are used. Therefore, both the observer at the station and his counterpart in the administration who receives and processes the telegrams containing hydrometeorological observation data must have perfect knowledge of the codes for transmitting observation data.

The administrations of the hydrometeorological service receiving the telegrams containing hydrometeorological observation data for each period consolidate them into combined telegrams and once again transmit them by radio or send them by telegraph to rigidly established addresses. These combined telegrams are sent to neighboring administrations of the hydrometeorological service, ^tThe Hydro-~~met~~ meteorological bureau, Central Institute of Forecasting and other organizations interested in receiving detailed information on the condition of the sea.

Code KN-02 serves to transmit the observations of shore hydrometeorological stations and posts.

Code KN-02¹ permits reception of the observations from

stations (See "The Code of Shore Hydrological Observations on Marine Stations and Posts KN-02". Gidrometecizdat L. 1949) in great volume. But to receive high quality messages with the aid of this code requires careful work on the part of the observer encoding the results of his observations and of his colleague who must decode the message. One of the peculiarities of this code is that it gives the observer an opportunity to transmit in an economical and concise form the most essential data from all his observations. For example, the ninth group of the code requires of the observer the ability always to sort out the most important and the most characteristic phenomena which occurred in the ice coating during the time since the last observation of the ice. It is very important that the observer promptly communicate, using this group, when complete freezing, opening, etc. took place.

The numbering of the separate groups allows us to use the code expediently, as each extra word (group) in a telegram creates an excessive expenditure of government funds, an unnecessary load on the telegraph, takes away spare time from the observer and the technician who decodes the telegram. In connection with this, the administrations of the hydrometeorological service carefully examine the entire network of marine hydrometeorological stations and establish separately for each station the quotas of telegrams for each period. For example, it is not always expedient to use the third group (in which data on disturbances is given) for a night period of observation. It may happen that a station generally may not use the third group if it is so located that its observations of disturbances are either impossible or unrepresentative. (not representative
for the area in which the station is located.)

It may also happen that observations of the tides are not always possible or expedient to transmit in telegrams.

The telegrams received undergo careful processing, which consists of decoding, entering the observations into special journals and transferring the results of the observations to hydrosynoptic charts. Prior to consolidating the separate telegrams they must pass a careful and critical examination and all the inaccuracies resulting from telegraphic errors must be removed. It is desirable to make the critical examination at the same time that data is transferred from the telegrams to hydrosynoptic charts. Such an examination proves to be extremely practical, since the comparison of the observations of the different stations becomes easier in the transfer to the charts.

In transferring the observations to the charts we use standard symbols (indicated on Figure 1). These same standard symbols are used in transferring the data from hydrometeorological observations to hydrosynoptic charts. The hydrometeorological observations of ships are also transmitted by radio in code. For this we use code KN-09. (The code for composing synoptic telegrams on board ships KN-09. Gidrometeoizdat, L., 1949) The hydrosynoptic charts are carefully analyzed. [See Figure 1 on following page]

Before making an analysis of the hydrosynoptic chart or a critical examination of the telegraph messages of the various stations, it is necessary to acquaint oneself carefully with the reports of the stations. It is highly desirable that the worker making the analysis of the hydrosynoptic chart acquaint himself with observation conditions at the stations not only on the basis of the written reports, the com-

FIGURE I

STANDARD SYMBOLS USED IN CONSTRUCTING HYDROSYNOPTIC CHARTS

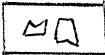
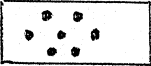
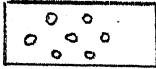
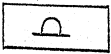
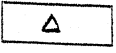
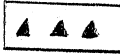
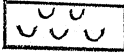
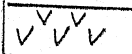
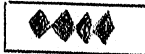
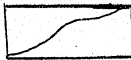
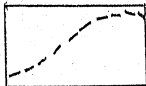
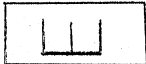
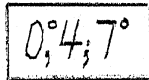
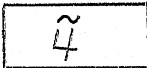
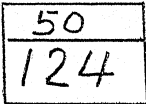
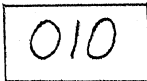
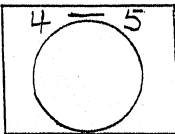
No P/P	Standard symbol	Meaning of the symbol
[1]	[2]	³ [4]
1		slush, soft snow, sludge ice
2		ice rind, thin ice, young ice
3		smooth ice fields
4		segments of smooth fields
5		coarsely chopped ice
6		finely chopped ice
7		ice pulp
8		floeberg
9		stationary ice, (shore ice, land floe, foot of a land floe), smooth

FIGURE I (cont'd)

[1]	[2]	[3]
10		stationary ice (shore ice, land floe, foot of a land floe), smooth, with cracks in various directions
11		stationary ice (shore ice, land floe, foot of a land floe), smooth, with cracks in a definite direction
12		stamukh (ice)
13		iceberg
14		ice packs
15		snowdrifts
16		place where snow has thawed out, or washout
17		tidal lead or unfrozen patch of water in ice
18		watery sky
19		icy sky
20		density of the floating ice in balls

[1]	[2]	[3] FIGURE I (cont'd)
21		degree of packing of the ice in balls
22		degree of the break up of the ice in balls
23		direction of drift
24		ice cutting, rotating
25		compression of the ice
26		thinning of the ice
27		channel in stationary ice
28		passable road on stationary ice
29		pedestrian road on stationary ice
30		thickness of the ice in centimeters.
31		water on the ice

[1]	[2]	[3] FIGURE I (Cont'd)
32		rim of ice or edge of ice of detectable thickness
33		supposed rim or edge of ice of detectable thickness
34		force and direction of wind: long tail -- 2 balls; shore -- 1 ball
35		calm
36		temperature of the water in degrees
37		balls of disturbance
38		visibility seaward (in cable lengths), sea level above zero of the post (in centimeters)
39		atmospheric pressure in millibars
40		current: long tail -- half knot; short -- a quarter knot
41		width of the stationary ice in kilometers

Note: If the ice fields, segments of the fields, and land floes are not smooth but packed with snowdrifts, with places that have thawed out, with washouts, then the standard bymbols of these forms are supplemented with the appropriate symbol (Numbers 15, 15, 16).

ments of inspectors or other persons who have visited the stations, but for him to have been at most, or better yet, all the hydrometeorological stations.

While analyzing the hydrosynoptic chart it is necessary to examine carefully the general synoptic condition over the sea not only for the current period of observation, but also for the previous one or two periods. For this purpose, the regular synoptic charts analyzed by the weather service personnel are used.

The resulting analysis of the hydrosynoptic charts should be an accurate presentation of the condition of the sea not at the particular station, but in that area of the sea. At the same time we must not divide up the sea into artificial areas, for example, into areas for which summaries are written, or into areas for which storm warnings are given. The areas must be made up naturally. We should divide out areas in which there is a rise or fall in the tides, an increase or a decrease in the disturbance, a fall or rise in the temperature of the water, a compression or thinning of the ice. Naturally, such a division is made only when the chart analyst can clearly visualize these physical processes developing in the sea.

Naturally, it is simpler to make an analysis of the hydrosynoptic charts when more observations are placed on the chart. The shipboard hydrometeorological observations play an especially large role. They are important first of all because in most cases they are free of the distortions caused by the shore line. Particularly the shipboard and sometimes the island observations facilitate following the development of this or that phenomenon of the sea. The development of a disturbance at sea is most difficult to follow in coastal

observations. That is why data on the actual condition of the sea can be most valuable only when the hydrosynoptic charts contain a sufficient number of shipboard observations.

After an analysis of the hydrosynoptic chart is made, when an account is drawn up of the processes developing at sea, the summary is made up. The summary, depending on which purpose it is to serve, can be made for several conventional areas or for natural areas. The summary can be made simply as a description of the data or as an examination of the phenomena of the sea, in their development and casual relationships. The summaries made as by the second method are the most advantageous. In no case, however, in making summaries can we replace false observations by deficient ones. It is necessary to remember always that a summary is a document, drawing an actual map of the phenomena for the time of observations. In the summaries, depending ^{ON} ~~ON~~ their intended purpose, the results of indirect observations of various stations are also included. For this the more representative stations are selected, whose observations are characteristic for entire sea areas. However, one need not always use the observed data of representative stations only. Frequently, it is important for commercial organizations to know the condition of the sea in any port ~~or~~ mouth of a river, harbor, etc. Since the compilation of the summaries is not just for one purpose, before commencing it, it is necessary to clarify which regions and exactly what information concerns the organizations for which the summaries are intended.

4. THE SIGNIFICANCE OF WEATHER FORECASTING AND THE CONDITION OF THE SEA

It would be simplest to trace the value of forecasting to

the national economy according to the problems that are ordinarily dealt with in the organs of the hydrometeorological service.

Example 1.

The shiphandler, who must tow a river barge from the mouth of the Danube to the mouth of the Dnepr, cannot be satisfied with the information that at the time when he is putting to sea this entire route would be free of heavy seas. He also is not satisfied with information on the mean long term values of the frequency of rough seas. All this information does not preclude the possibility of the development of heavy, perhaps even stormy seas or of heavy swells enroute. To get to the point, the shiphandler is interested only in one question: whether, during the time required to transfer the barges, there will be a disturbance greater than 4 balls, since the barge cannot withstand seas heavier than that, or, in any case, would encounter great difficulties enroute if the sea was rough.

To ensure the successful transfer of the barge by the shiphandler it is necessary to assist him in selecting a time when, on the basis of the analysis of the hydrometeorological situation, he will be assured with some degree of accuracy that there will be no strong disturbances enroute.

Example 2.

During navigable periods large quantities of petroleum are transported from Baku to Astrakhan' and further upstream along the Volga. The Volga-Caspian Canal is so shallow that sea-going tankers cannot reach Astrakhan'. They must stop near the roads and transfer

the oil to barges, which ship between the area of the roads and Astrakhan'. The area of the Volga-Caspian Canal is subjected to considerable ebb-flood tide variations up to 120 centimeters. The trip from the roads to the end of the shallow part of the canal takes about a day. From the time of filling the barge with oil to the time of passing the shallow part of the canal, the level of the sea may change as much as 120 centimeters. If while filling the barge, one calculates on the basis of the mean level, then with a full oil load the barge may find itself aground in case of an ebb tide, and in case of a flood tide it will make a long trip not completely loaded. Therefore, the proper functioning of the tanker fleet requires short-term forecasts of the ebb-flood tide variations on the Volga-Caspian Canal.

Example 3.

When planning cargo transfer from the Caspian Sea to the port of Astrakhan', and also during the fall fishing season in the northwestern part of the Caspian, the question arises: when will ice appear in the Astrakhan' area? The average date for the appearance of ice in this area is the beginning of December, but frequently the ice appears in November. At the same time there are cases on record when ice appeared in January. Hence, the period in which the ice may appear ~~extends~~ over many days. If it is taken into account that in the course of one day of the fall fishing season 30,000 centners of fish can be caught and large quantities of cargo can be transported, the necessity of forecasting long in advance the appearance of the ice and the end of the navigable period becomes apparent.

Example 4

As a result of a storm a vessel on a certain course on the open sea suffered a wreck. The main engines or the steering gear and the radio station went out of order. For a successful solution of the question of what area should be searched for the disabled vessel, it is imperative to have a forecast of its drift, including also the direction of drift.

Example 5

For the past few years (beginning in 1928) the level of the Caspian Sea has dropped more than 2 meters. In view of the shallowness of the northern part of the Caspian such a drop caused large areas of the sea to dry up.

As a result, several fishing establishments built on the beach were tens of kilometers from the waterline. The approaches to the ports and the ports themselves grew shallow. Tremendous facilities were required for dredging in order to maintain the ports in a navigable condition. Suffice it to note that a drop in the mean annual level of 5 centimeters requires dredging in the Volga-Caspian Canal costing about one million rubles. In order to plan this work it is necessary to have forecasts of variations in the water level for a period not less than 1 year in advance indicating also the mean level for each month.

The calculations which are made for hydrotechnical construction projects and generally for major construction on the shores of the Caspian require forecasts of the sea level for a long period ahead --

say for a few years, inasmuch as without these forecasts the major construction projects may find themselves in the position of the fishing establishments which are left tens of kilometers from the waterline.

The examples described show conclusively just how important hydrometeorological forecasts are in solving many problems of the national economy. The data on current weather and sea conditions and the mean long-range characterizations of the hydrometeorological regime play only an auxiliary role. The forecasts fill in the deficiencies of other hydrometeorological information -- which gives the probable condition of the sea in the immediate future. However, in many instances it becomes necessary to utilize the mean and extreme characteristics of the maritime regime, based on long-range observations, because of the impossibility of giving a sufficiently well founded forecast.

CHAPTER III

SOME GENERAL CONCEPTS

5. DIVIDING FORECASTS INTO REPORT PERIODS

Forecasts released by organs of the hydrometeorological service are divided into proper periods, that is, into periods separating the moment of compilation of the forecast from the last effective moment of the forecast in the following main groups:

- (1) short-term -- report period not over 2 days;
- (2) three-day-report period not over 3 days;
- (3) long-range forecasts of a short report period --
report period up to 1 month;
- (4) long-range -- report period from 1 to 2 months;
- (5) seasonal -- report period up to 6 months;
- (6) yearly -- report period up to 1 year;
- (7) extremely long-range -- report period over 1 year.

These divisions are provisioned and are fixed by methods accepted by the hydrometeorological service. Since these methods are far from perfect and uniform it is natural that every forecast of the same ^{PHENOMENON} ~~phenomenon~~ with a shorter reporting period had best be released only when one can count on the realization of the predicted phenomenon with a greater degree of probability than with a forecasting method having a longer reporting period. In deciding on the best method of forecasting, the reliability of the method is considered.

By the serviceability of a method is determined the feasibility

of employing one or the other method in establishing regularly scheduled forecasts. The reliability of the method is figured according to the products of previous observations by comparing predicted values with those actually observed.

For an objective qualitative comparison of methods having different report periods intended to forecast various phenomena in different seas or regions of seas, the errors in forecasting are expressed in percentages of amplitude variations in the predicted phenomena.

6. LONG TERM AMPLITUDE

The difference between the greatest and least values of a predicted phenomenon observed under concrete physiogeographical conditions for a definite period of time is conventionally called an amplitude (In practice, to determine the long-term amplitude the series of observations for all years is used.)

The long-term or calculated amplitude is examined in relation to the predicted phenomenon and the report period of the forecast.

The swing from the greatest to the least (inclusive) values of any phenomenon determined by many observations made in the course of many years is called a long term amplitude. For example, if the earliest appearance of ice at a given point was noted on 8 November, and the latest appearance of ice was observed on 18 January, then the amplitude in this case is 72 days. The amplitude of the same phenomenon for different areas varies within extremely wide limits. The long-term amplitude for periods of complete freezing for Nikolayev is only 54 days, while for Ochakov it is 93 days.

At the same time, the amplitude for periods of freezing at many points scattered in the estuary region of the Amur River varies within the limits of from 27 to 30 days. The amplitudes for periods of this phenomenon in the estuary of the Pechova River are also not great -- only about 45 days.

The long-term amplitude of other phenomena is determined similarly. For example, if we are to predict a monthly mean sea level, the long-term amplitude of variations in the level will be the difference between the highest and lowest levels observed in the month in question.

Where the observations of a predicted phenomenon are not high, it becomes necessary when determining the long-term amplitude, to establish whether or not the observations during years having extreme values for the phenomena are being considered.

This can be established by analyzing the long-term observations at adjacent points where similar observations were made but over a longer period of time. If as a result of the analysis it is established that the years having extreme values in the observed series of observations are missing, it is necessary to attempt to interpolate the missing extreme values. To do this we construct a graph showing the relationship between all the values of the observation series interesting us and the observation series for a large number of years. If the relationship turns out to be close then, using the extreme years of the series in the area with a large number of observations, and the graph, the probable amplitude values for the area with few observations are established. If we fail in bringing the amplitude of the short observation series to the large one in this

manner, due to a poor relationship, then it is customary to take the amplitude calculated from a short-term series, if, despite the short observation series, the extreme values of the observed element enter into it, the transfer to the large series of observations is not undertaken.

To evaluate the reliability of the forecasting method a full long-term amplitude for periods of phenomenon appearance can be accepted only if the forecast for the report period is released up to the time when, judging by the earliest date, the predicted phenomenon can become a reality. In the contrary case, when the release date of the forecast falls between the earliest and latest periods in which this phenomenon is observed, the long-term amplitude must be decreased to a corresponding value. For example, the earliest appearance of the ice is observed on 11 November, the latest one -- 20 January, while the method allows the forecast to be released only on 1 December. Naturally, in evaluating this method the amplitude must be decreased from 71 to 51 days, as this method excludes the possibility of giving a forecast of the appearance of ice from 11 November to 1 December.

7. CALCULATED AMPLITUDE

The swing from the highest to the lowest value which a predicted phenomenon can have for the effective time of the forecast is called a calculated amplitude. In order to explain more easily the difference between a long-term and a calculated amplitude, several examples are given here, the examination of which will clarify the necessity of resorting in some cases to a calculated, instead of a long-term amplitude. Let us suppose that it is necessary to fore-

cast the mean monthly level at one of the points in the Caspian Sea. It is well known that in the course of many years the level of the Caspian undergoes very large variations. A long-term amplitude is several meters while the month to month changes in the level are always only a few centimeters. Therefore, it is inexpedient to use the long-term amplitude in evaluating forecasts.

The correct evaluation of such forecasts, requires use of the calculated amplitude which represents the swing in the mean monthly level changes at a given point from month to month for a number of preceding years.

For example, let us suppose we are required to make a forecast of disturbance in balls at some point at sea for 12 hours in advance. At this point disturbances of from 0 to 9 balls have been observed, that is, the disturbance amplitude is 9 balls. However, in our calculations we cannot use this amplitude, because from an analysis of the products of previous observations it becomes clear that for 12 hours no disturbance has ever changed more than 6 balls. Therefore, in formulating disturbance forecasts for 12 hours it is necessary to use a calculated amplitude, the change in the degree of disturbance being 6 balls instead of 9. A calculated amplitude is determined by calculating from the products of previous observations the differences between the values which the predicted phenomenon gave for an increment of time equal to the effective period of forecasts. In calculating the differences we establish the possible positive and negative deviations in the progress of the predicted phenomenon, that is, the rise and fall in the water temperature, the strengthening or weakening in the disturbance, etc., in comparison with the time of releasing the forecast. The absolute values (without

the plus or minus sign) of the greatest positive and negative deviations are collected and the result is used as the amplitude. Practice has shown that the general series of observations, from which the amplitude is determined, must contain not less than 50 cases.

Table 1 gives examples of computing the calculated amplitude for predicting mean daily values of water surface temperatures and mean daily figures on the sea level for a forecast report period of one day.

TABLE I

Dates	Mean daily water temperatures	Daily temper- ature change	Mean Sea level	Daily change in level
[1]	[2]	[3]	[4]	[5]
1	25.4		114.8	
2	25.9	0.5	116.5	1.7
3	25.6	-0.3	130.2	13.7
4	24.7	-0.9	120.8	-9.4
5	25.5	0.8	120.2	-0.6
6	26.2	0.7	119.0	-1.2
7	26.4	0.2	118.8	-0.2
8	26.1	-0.3	114.5	-4.3
9	26.2	0.1	116.0	1.5
10	25.5	-0.7	121.5	5.5
11	25.2	-0.3	130.2	8.7

TABLE I (cont'd)

[1]	[2]	[3]	[4]	[5]
12	23.0	-2.2	202.2	72.0
13	21.2	-1.8	189.0	-13.2
14	20.8	-0.4	152.2	-36.8
15	21.2	0.4	122.2	-30.0
16	21.7	0.5	117.2	-5.0
17	22.4	0.7	121.0	3.8
18	22.4	0.0	121.0	0.0
19	22.7	0.3	118.8	-2.2
20	22.9	0.2	126.8	8.0
21	22.2	-0.7	138.2	11.4
22	21.4	-0.8	147.0	8.8
23	21.0	-0.4	140.5	-6.5
24	20.3	-0.7	157.8	17.3
25	20.4	0.1	130.0	-27.8
26	21.9	1.5	114.8	-15.2
27	22.0	0.1	120.8	6.0
28	20.6	-1.4	157.2	36.4
29	19.4	-1.2	151.5	-5.7
30	19.0	-0.4	125.0	-26.5
31	19.5	0.5	108.5	-16.5

In these series of observations the greatest daily drop in water temperature reached 2.2 degrees, and the rise did not exceed 1.5 degrees. Therefore, the calculated amplitude can be considered as the sum $2.2 + 1.5 = 3.7$ degrees.

The greatest daily increase in level was 72.0 centimeters, and the decrease was 36.8 centimeters. The calculated amplitude in this case will be 108.8 centimeters (Obviously, the calculated amplitude for various seasons may be found to vary; therefore, it should be determined separately for each season, and sometimes for each month).

In order to save space only 31 values of water temperature and level were examined in the examples in Table 1. Usually, though, in determining the calculated amplitude we should take not less than 50 observations.

Sometimes the forecasts are of interest to economic organizations only on condition that the allowable error in forecasting does not exceed some of the previously given values independent of the long-term or calculated amplitudes. In these cases the method is evaluated according to what percentage of the general number of forecasts are the forecasts with errors not exceeding a given value. Usually, only those methods are used in operational work which guarantee that in 80 percent of the forecast data errors do not exceed given limits.

8. THE RELIABILITY OF A FORECAST PROCEDURE

The probability or frequency of errors in forecasting is the criterion that serves for the evaluation of a forecast procedure.

If we determine the frequency of errors in forecasts using a definite procedure, then we also determine the reliability of this forecast procedure.

In calculating the errors in forecasting in absolute values (days, centimeters, degrees, etc.) we lose the possibility of comparing the reliability of the same procedure as it is applied to various geographical areas, since the amplitude of predicted phenomena is not the same. That is why the reliability of a procedure is usually figured in percentages of the long-term or calculated amplitude. The long-term or calculated amplitude gives us the opportunity to approach the evaluation of the forecast procedure objectively enough. Most frequently the largest allowable error is established at a quantity equal to 20 percent of the long-range or calculated amplitude. As a rule, errors in 80 percent of the forecasts do not exceed 20 percent of the long-term or calculated amplitude when the coefficient of correlation, characterizing the continuity on which the forecast is based, is not less than ± 0.75 , ± 0.80 .

The coefficient of correlation, usually designated by the letter r , is a measure of the quantitative cohesion in the case of a linear function. It represents an abstract number falling within the limits of from -1 to $+1$. In cases of functional relations the coefficient of correlation equals $+1$ or -1 . In the first case ($r = +1$) the dependent variable (function) increases (or decreases) with an increase (or decrease) of the independent variable (argument). In the second case ($r = -1$) the function decreases (or increases) with an increase (or decrease) of the argument. In the complete absence of continuity the coefficient of correlation $r = 0$. When the value of

r changes from 0 to 1, the correlation is called positive, and from 0 to -1, the correlation is called negative.

An example of calculating the coefficient of correlation is shown in Table 2.

A comparison of the forecast reliability with a so-called natural reliability serves as a criterion of the effectiveness of the process. It is useful to construct a diagram showing the distribution of the deviations from the amplitude in percentages, in order to clarify the effectiveness of the process. In constructing the diagram we usually make the calculations shown in Table 2. Verified forecasts are made by the continuity which is the basis of the procedure or by other means. The data for past periods serve as the original materials for forecasting. 25 to 50 verified forecasts must be made for an evaluation of a procedure. For this it is most desirable that these forecasts not be compiled on the basis of those observations which served in establishing the evaluated function. However, because of insufficient observations, especially in working out seasonal forecasts, it is permissible to take data used in working out a procedure. Having received a sufficient number of forecasts, the absolute errors in forecasts are calculated by seeking the differences between the predicted and the actual values. Next, the long-term or calculated amplitude is determined, and then the values which are obtained from the amplitude, 5, 10, 15, 20, 25, 30, 35 percent, etc. All the errors in forecasts are split up into groups in an ascending order of 5 percent of the amplitude: 0 - 5, 0 - 10, 0 - 20, 0 - 25 etc. We then calculate the number of forecasts appearing with errors falling into each of these groups.

Then we calculate what percent of the general number of forecast data comprises the number of forecasts falling into each group. Having received all the enumerated data, an error-distribution curve on verified forecasts, made by an appropriate procedure (Figure 2), is made on millimeter graph paper. To do this, the forecast errors are plotted along the ordinate axis, on a scale of 1 centimeter = 5 percent, in percentages of the amplitude, and the reliability along the abscissa axis in percentages of the general number of cases of forecasts having some error. In other words, the calculations made in this manner allow us to read on the diagram along the abscissa from zero consecutively by 5 or 10 percent, and to place dots at the intersections of the corresponding ordinates through which the reliability curve is then drawn. The dots made in this manner on the diagram are joined in a smooth curve, which is called a distribution curve in percentages of error amplitudes of verified forecasts made by the procedure. [See Figure 2 on following page]

On this same graph a distribution curve of deviation from the norm is drawn in percentages of the amplitude of that phenomenon whose forecast procedure is being tested. All calculations from this are made entirely in an analogous manner, only in this case it is not the predicted and observed values that are compared, but the mean, calculated by examining many observations, and the observed, that is, each number of the series.

Both curves are shown on Figure 2, and all the calculations necessary in constructing them in Table 2. Examining this figure it follows that the closer the curve is to the abscissa axis,

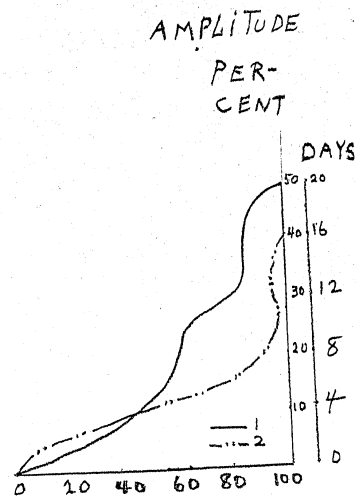


Figure 2. Reliability curves. 1 - natural reliability, 2 - reliability of the procedure

the less is the procedure error in percentage of the amplitude; but the larger the effectiveness of the procedure, the larger the area between the procedure reliability curve and the natural reliability curve. [See Table 2 and equations on following pages]

Explanation of Table 2.

In column 1, Table 2, we have numbers in sequence; in 2, the years for which the observations were made; in 3, the independent variables, for example the water temperature, indicated by the letter x;

TABLE 2

No P/P	Year	Independent variable x	Dependent variable y	$\Delta x = x - \bar{x}$	$\Delta y = y - \bar{y}$	Δx^2	Δy^2	$\Delta x \Delta y$	$(\Delta x + \Delta y)$
1	2	3	4	5	6	7	8	9	10
1	1915	4.0	19	0.5	-4	0.25	16	-2.0	3.5
2	1916	4.7	21	1.2	-2	1.44	4	-2.4	0.8
3	1917	0.0	37	-3.5	14	12.25	196	-49.0	10.5
4	1918	0.6	35	-2.9	12	8.41	144	-34.8	9.1
5	1919	5.0	9	1.5	-14	2.25	196	-21.0	12.5
6	1920	3.4	28	-0.1	5	0.01	25	-0.5	4.9
7	1921	2.4	33	-1.1	10	1.21	100	-11.0	8.9
8	1922	2.5	34	-1.0	11	1.00	121	-11.0	10.0
9	1923	4.8	12	1.3	-11	1.69	121	-14.3	9.7
10	1924	3.4	26	-0.1	3	0.01	9	-0.3	2.9
11	1925	4.2	4	0.7	-19	0.49	361	-13.3	18.3
12	1926	4.4	22	0.9	-1	0.81	1	-0.9	0.1

Table 2 (cont'd)

[1]	[2]	[3]	[4]	[5]	[6]	[7]	[8]	[9]	[10]
13	1927	3.9	17	0.4	-6	0.16	36	-2.4	5.6
14	1928	3.5	20	0.0	-3	0.00	9	0.0	3.0
15	1929	3.0	26	-0.5	3	0.25	9	-1.5	2.5
16	1930	4.5	24	1.0	1	1.00	1	1.0	2.0
17	1931	1.4	42	-2.1	19	4.41	361	-39.9	16.9
18	1932	2.4	29	-1.1	6	1.21	36	-6.6	4.9
19	1933	6.0	3	2.5	-20	6.25	400	-50.0	17.5
20	1934	2.8	31	-0.7	8	0.49	64	-5.6	7.3
21	1935	4.7	21	1.2	-2	1.44	4	-2.4	0.3
22	1936	5.4	21	1.9	-2	3.61	4	-3.8	0.1
23	1937	2.9	24	-0.6	1	0.36	1	-0.6	0.4
Σ		79.9	538	-	-	49.00	2219	-272.3	-
n		23	23						
Average		$\bar{x}$ 3.5	$\bar{y}$ 23						

TABLE 2 (Cont'd)

$(\Delta x + \Delta y)^2$	y_{ϕ}	y_B	Error	Error	Scale		Frequency in	Frequency in
			$y_{\phi} - y_B$	$y_{\phi} - y_B$	0/0	ДНН	Δy	Δy
			14	15	16	17	18	19
11	12	13						
12.25	19	26.3	-1.3	-1	0-5	2.0	13	26
6.64	21	14.5	4.5	4	0-10	3.9	43	43
110.25	37	44.3	-5.3	-5	0-15	5.8	78	57
82.81	35	39.0	4.0	4	0-20	7.8	91	61
156.25	9	14.8	5.8	6	0-25	9.8	96	65
24.01	28	23.6	4.4	4	0-30	11.7	96	78
79.21	33	29.1	3.9	4	0-35	13.6	96	87
100.0	34	28.5	5.5	6	0-40	15.6	100	87
94.09	12	15.9	-3.9	-4	0-45	17.6		87
8.41	26	23.6	-2.4	-2	0-50	19.5		100
334.89	4	19.2	-15.2	-15	0-55	21.4		
0.01	22	18.1	3.9	4	0-60	23.4		

TABLE 2 (Cont'd)

[11]	[12]	[13]	[14]	[15]	[16]	[17]	[18]	[19]
31.36	17	20.9	-3.9	-4	0-65	25.4		
9.00	20	23.1	-3.1	-3	0-70	27.3		
6.25	26	25.8	0.2	0	0-75	29.2		
4.00	24	17.5	6.5	6	0.80	31.2		
285.61	42	34.8	7.2	7	0.85	33.2		
24.01	29	29.1	-0.1	0	0-90	35.1		
306.25	3	9.3	-6.3	-6	0-95	37.0		
53.29	31	26.9	4.1	4	0-100	39.0		
0.64	21	16.5	4.5	4				
0.01	21	12.6	6.4	8				
0.16	24	26.3	-2.3	-2				
1723.40								

TABLE 2 (Cont'd)

[Equations shown at bottom of Table 2]

$$\sigma_x = \sqrt{\frac{49.00}{23}} = 1.46;$$

$$\sigma_y = \sqrt{\frac{2219}{23}} = 9.82;$$

$$r = \frac{-272.3}{1.46 \cdot 9.82 \cdot 23} = \frac{-272.3}{329.76} = -0.82;$$

$$(y - \bar{y}) = \frac{\sigma_y}{\sigma_x} r (x - \bar{x});$$

$$(y - 23) = \frac{9.82}{1.46} - 0.82 (x - 3.5);$$

$$y - 23 = -5.51x + 19.28;$$

$$y = -5.51x + 42.28;$$

$$y = 42.3 - 5.5x.$$

in 4, the dependent variable $\bar{y}$. This can be a predicted value, for example the period of appearance of ice expressed in days and figured from the beginning of any month until the appearance of ice. In column 5 we write the differences $\Delta x = x - \bar{x}$, whereupon $\bar{x} = \frac{\sum_1^n x}{n}$, where n is the number of lines or the last index number; $\sum$ represents the sum. In other words, $\bar{x}$ is the mean of all the argument values listed in column 3.

In column 6 we write the differences $\Delta y = y - \bar{y}$; $\bar{y}$ is calculated the same as $\bar{x}$. In columns 7 and 8 we have the values Δx and Δy squared; in 9, we have the product of the values listed in column 5 and 6.

Column 10 and 11 are made only for the purpose of checking the preceding calculations; in 10 we have the sums of the values listed in columns 5 and 6 without considering the mathematical sign; in 11, the same sum squared. The basis of the check is the well known formula:

$$\sum_1^n (\Delta x + \Delta y)^2 = \sum_1^n \Delta x^2 + \sum_1^n \Delta y^2 + 2 \sum_1^n \Delta x \Delta y.$$

All these values are in the table. Therefore, it is easy to make the check having compared the value in column 11 on the line marked

with the sum of the values on the same line in columns 7, 8, 9; thereupon, the value in column 9 must be first multiplied by 2. The equality of these figures shows that the calculations were made properly, and an inequality indicated that an error has crept into the calculations.

After establishing that the calculations have been made correctly the quadratic deviations are calculated by the formulas:

$$\sigma_x = \sqrt{\frac{\sum \Delta x^2}{n}} \quad \text{AND} \quad \sigma_y = \sqrt{\frac{\sum \Delta y^2}{n}}.$$

Having calculated the quadratic deviations it is easy to find the coefficient of correlation:

$$r = \frac{\sum \Delta x \Delta y}{n \sigma_x \sigma_y}.$$

After this we can construct the regression equation using the formula

$$y - \bar{y} = \frac{\sigma_y}{\sigma_x} r (x - \bar{x}).$$

All these calculations are brought out in Table 2.

Calculation of the reliability of the equation is shown in columns 12, 13, 14, 15.

In column 12 the data from column 4 is rewritten for convenience.

In column 13 we have the data obtained by substituting the values from column 3 in the regression equation and associated calculations.

In column 14 we write the differences in the figures in column 12 and 13. These differences represent nothing else than the errors in the equation.

In column 15 those same errors are presented, rounded off to whole numbers.

Further, the long-term amplitude of a phenomenon is calculated. To do this we take the largest and smallest figures out of column 4 and find the difference between these figures. In this case the largest figure is 42 days and the least 3 days. The amplitude is 39 days. 20 percent of the amplitude is 7.8 days.

In column 16 we have the percentage scale. In column 17 we have the same scale expressed in days.

In column 18 we have the percentage of error frequency shown in column 15, falling into appropriate gradations, written in column 17.

In column 19 the frequency percentage observed in the deviation Δy listed in column 6 along the same gradation of column 17, is calculated.

The frequency percentage indicated in column 18 is called a reliability equation or the reliability of the forecast procedure, if the regression equation allows us to make forecasts.

The frequency percentage indicated in column 19, is called a natural reliability.

In this manner, the probability or the frequency of the deviation in the observed values of one or the other element of the regime from its mean value is called a natural reliability, or, in other words, the frequency of deviation from the norm.

Actually, when these curves come together or almost come together, application of the procedure has no effect when comparing with the mean long term value, that is, the errors will be just as great when the forecast procedure will be used as when instead of the forecast, we designate the mean long-range value each time.

By comparing the reliability of a forecast procedure with the natural reliability, a comparison is made of the reliability of a procedure with the reliability of the so-called inertia forecast. A forecast is designated as inertia when it is expected that the condition of one or the other element in the regime, observed at the time of formulating the forecast, will be preserved unchanged in the future. For example, the ice condition at sea in January was 20 percent above normal; in making an inertia forecast for February, we assume that the ice condition at sea in February will likewise be 20 percent above normal. Or, another example: the water temperature today is 15 degrees, we expect the water temperature to be 15 degrees tomorrow too. Having made a sufficient number of inertia forecasts in this manner, we construct an error-distribution curve for inertia forecasts by a method very similar to the above indicated method and compare it with the error-distribution curve for the employed procedure of forecasting. Quite frequently the reliability of the inertia forecast is higher than the natural reliability. To evaluate forecast procedures it is necessary to take for comparison the error-distribution curve for inertia forecasts in those cases when the distribution curve is constructed as a percentage of the calculated amplitude, but when the long-term amplitude is taken as a basis of calculation the reliability of a procedure is compared with the natural reliability.

11

The justification of forecasts is evaluated in the same manner as the reliability of a method. The difference in this case is that the forecasts are not given from the products of past years, but are evaluated by the actual predictions of one or the other element in the field. Usually, in forecasts we do not give exactly that value which develops as a result of an objective application of a method, but we take some interval whose mean value is the figure obtained by an appropriate method. This is done because the relations on which the forecasts are made are not functional, but have only a certain degree of reliability. This interval is most acceptable when it is not over 20 percent of the long-term or calculated amplitude, avoiding, of course, the fractional values in setting the interval limits of the expected phenomenon. In evaluating the justification of forecasts, we take the mean value of the predicted figure. This figure also serves in calculating errors by comparing it with the observed. The absolute error obtained in this manner is expressed in percentage of amplitude.

It is usually accepted to consider the error not over 5 percent of the amplitude as being responsible for perfect forecasts. Forecasts with errors of not over 10 percent of the amplitude are considered good, and with errors of not over 20 percent as satisfactory.

Where forecasts are made simultaneously for various areas of the same sea or when forecasts are made for several seas, the general evaluation of the forecast is made by determining the number of perfect, good, satisfactory and unsatisfactory forecasts. If it appears that of the general number of forecast areas the number of

unsatisfactory forecasts is more than 20 percent, then the entire forecast is considered unsatisfactory.

Inasmuch as forecasts with long report periods usually have a smaller reliability than forecasts with a short report period, it is best to carry on a parallel development of a method having a lesser report period but a large reliability in developing a long-term forecasting procedure. It is best to construct this work for long-term forecasting, because it is necessary to be able to control the errors in the basic long term forecast and when necessary to issue corrections to this forecast.

We generally consider that any forecast having a short report period, that is, having a later date of release, is closer to reality and, therefore, appearing in the light of the new forecast, it replaces the earlier released forecast of the same phenomenon. Although this situation is obvious, a proper organization of the forecasting service requires that each forecast modifying an earlier released long-term forecast^t be specially marked: exactly which forecast is replaced by releasing the modified one. A modified forecast must be distributed to exactly the same addresses that received the original forecast. Failure to carry out this simple regulation may cause an organization receiving forecasts to become disorientated by receiving several forecasts of the same phenomenon.

CHAPTER IV

SHORT-TERM FORECASTS OF TIDES AND CURRENTS9. GENERAL SITUATIONS APPLIED IN ANALYZING
THE PATTERN OF TIDES AND CURRENTS AT SEA

A change in the pattern of currents produces a change in the tide pattern. The reverse situation is also true, that is, together with a change in the tide there is a change in the current. Bearing in mind the interrelation between the patterns of tides and currents it is best to examine them jointly. Variations in the sea tide appear as a summary characteristic of the influence on the water masses by a great number of factors. The forces creating flood tides determine periodic changes in tides and currents. The continental flow and sets occurring on the surface of the water together with evaporation determine the seasonal changes of these elements in the pattern. The changes in atmospheric pressure and the action of the wind over the surface of the sea bring about the seasonal and short-term changes in the currents and the variations in sea level.

The enumerated and other processes taken together are so complicated and varied that attempts at finding a general analytic expression for level height or speed and direction of flow in relation to the factors determining them have not even been undertaken. But such attempts, actually, are not necessary. The principle laid down back in 1775 by Laplas, of superimposing a small variation allowed us to develop further a method of harmonic analysis of tidal variations in sea level and currents. Using the harmonic analysis

method and processing the observation material for a previous period, we can precalculate the correct periodic variations in the tide and currents for a long period in advance. The principle of superimposing small variations brings us to the following: the duration of variations in the sea level brought about by the action of a periodic force is equal to the duration of this force. The effect of several forces acting simultaneously is collected and a general result of the action of all forces becomes the sum of the variations caused by each force taken separately. Therefore, the variations caused for various reasons can be examined independently.

This principle, effective in studying tide-ebb phenomena, can be applied successfully also in the empirical study of tide and current changes caused by other and, for that matter, nonperiodic influences.

It is not necessary to dwell here on the well developed method of harmonic analysis since much specialized literature has been devoted to it. Problems in seasonal and long-term changes in tides will be examined in Chapter VII which deals with long-term forecasting. In the present section only those short-term changes in tides and currents which are caused by wind action will be examined.

One of the forms of such tide and current variations is the sgonno-nagonniy (running-out - running-in). Usually in those areas where the flood tide is not great we pay attention to the sgonno-nagonnye phenomena and therefore, sgonno-nagonnye changes in tides and currents are released on high priority because of the practical

importance in navigation. The sgonno-nagonnye phenomena are important in shallow areas, where they are defined quite sharply, and it is just in these areas that there are frequent cases of insufficient water under the keels of moving ships. Only with these reasons can we explain that there are few indications in literature about the development of sgonno-nagonnye phenomena in seas with clearly defined flood tides or in such deep seas as the Black Sea. From this, of course, it does not follow at all that in these areas the sgonno-nagonnye phenomena are missing or are so small that they have no practical importance.

The mechanics of the sgonno-nagonnye phenomena is extremely complicated and, unfortunately, still little investigated. Some outlines of drift and gradient currents and the sgonno-nagonnye variations in the tide related thereto can be followed through, utilizing the results of the theory laid down by Ekman and developed subsequently by V. V. Shuleykin, V. B. Shtokman and other Soviet Scholars.

10. CHARACTERISTIC PECULIARITIES OF DRIFT AND GRADIENT CURRENTS ARISING FROM THE THEORY

When the depth of a sea is sufficiently great and the area in which we are examining the current is considerably removed from the shore, and when over most of the surface of the water the established wind (that is, it does not change with the passage of time) acts evenly in force and direction, then a current develops on the surface of the sea. This current deviates to the right of the wind direction at an angle of 45 degrees and at an absolute amount determined by the relationship;

$$U_0 = \frac{T}{\sqrt{2\mu\omega \sin \varphi}}$$

Here U_0 is the speed of the surface current: T is the force of the friction between the air and the surface of the sea; ω is the angular speed of the earth's rotation; φ is the geographical latitude of the place where the current is being studied; μ is the coefficient of internal turbulence friction.

The coefficient μ changes depending on the speed of the water motion, the depth, and the stratification of the sea. Until now the value of μ has not been reliably determined. The Table below may serve as a kind of orientation of the relation between the values of μ and the wind speeds.

Speed of wind in meters per second	3	5	7	10	20
μ in grams per centimeter seconds	28	110	220	430	1720

The current speed changes both in force and direction with increasing depth. The current speed decreases, and its direction deviates more and more to the right of the wind direction and finally at a certain depth the current moves in exactly the opposite direction in relation to the surface current (Figure 3). This depth is designated by D and is determined by the relationship:

$$D = \pi \sqrt{\frac{\mu}{\omega \sin \varphi}}$$

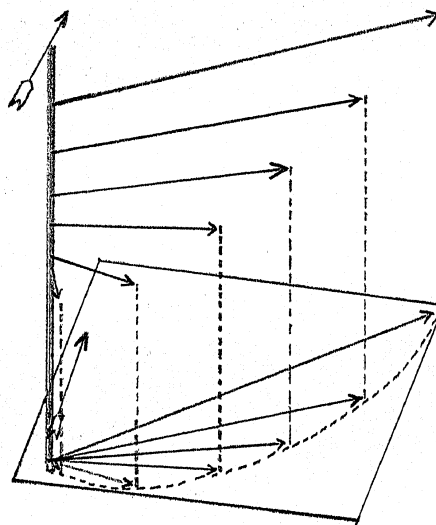


Figure 3. Diagram of the changes in current speed caused by
the wind in depth layers of friction D: 1 - wind, 2 - surface
current

The depth at which the current speed becomes directionally opposite is called the friction depth. The current speed at friction depth D is only $1/23$ of the speed created by the wind at the surface of the sea.

In the relationship determining the friction depth two variable figures are used: the coefficient of turbulence friction and the latitude of the location. But this indicates that the friction depth changes with the latitude of the location and with change in wind speed, as the coefficient of turbulence friction μ in turn changes depending on the wind speed. How the friction depth changes in relation to these Figures is shown in Table 3. [See Table 3 on following page]

In this manner, the friction depth decreases with an increase in the latitude of the location and increases with an increase in wind speed. Drift speeds on the horizons of the below mentioned friction depths are very small and we can disregard them.

If a sea has a depth H , a lesser friction depth, then the angle between the direction of the surface^e current and the wind decreases. For example:

H/D	0.50	0.25	0.10
angle between current direction and wind	45°	21.5°	0°

As was indicated above, the value of the friction coefficient μ changes with a change in the wind speed. Large wind speeds cause large current speeds, and this in turn develops large numbers of eddies in the water current. And this is the phenomenon which is

TABLE 3

Latitude in degrees	Changes in friction depth (meters) with changes in wind speed (meters per second)	
	10	20
[1]	[2]	[3]
45	90	180
50	87	175
55	85	170
60	82	165
70	80	160
80	75	150

50

calculated by the coefficient μ . That is why the current deviations from the wind are dependent not only on the sea depth, but also on the wind speed. In Table 4 we have the deviations of the current from the wind direction in degrees depending on the sea depth and wind force.

TABLE 4¹

Depth in meters	Wind in Balls								
	2	3	4	5	6	7	8	9	10
[1]	[2]	[3]	[4]	[5]	[6]	[7]	[8]	[9]	[10]
10	42	20	8	6	4	4	4	3	3
20	45	45	22	20	12	10	6	5	4
50	45	45	45	45	45	37	31	22	17
100	45	45	45	45	45	45	45	45	43
150	45	45	45	45	45	45	45	45	45

(Table calculated by V. A. Zenin but it has not been reduced here to a theoretical formula.)

Table 4 is calculated for the mean latitude of the Caspian Sea.

The change in the absolute values of current speeds from horizon to horizon occurs as in the case of a deep sea, that is, a gradual increase in speed is observed from the surface^e to the bottom. The law of increasing speeds changes depending on the sea depth.

If we make no suppositions about the vastness of the sea, as was done in the beginning of this section in examining drift, then in this case the presence of the shoreline will be substantially changed by the currents arising from the action of a wind. As a result of the wind action just over the surface of the sea a water current will arise which will unavoidably meet the shore in its path. The shore will act as a barrier, obstructing the free passage of the current. An accumulation of water brought by the current arising from wind action will take place at the shore. This will cause the surface of the water in the off-shore area to develop a slope and not remain horizontal. This will mean that the water pressure in the off-shore area and at some distance from the shore on a horizontal plane located at a particular depth will not be equal, that is, a pressure gradient will develop.

The pressure gradient will be a new force which causes the water to move in another direction. A new water current will arise which will flow perpendicularly toward the pressure gradient on its right. The entire water mass will move in such a way, that if the pressure increases from left to right in relation to the observer, then the flow of liquid will move directly from him. Naturally, the

pressure gradient runs perpendicularly to the shoreline. In connection with this the current caused by the pressure inequalities within the water will flow along the shore. If the level of the shore is much higher than the sea, the current will flow from left to right relative to an observer facing seaward. If the shore level is lower, however, than at some distance at sea removed from it, then the direction of the gradient current will be directly opposite to that of the first case, it will flow from right to left relative to an observer facing seaward.

This rather simple picture of a gradient current would be observed if the motion could be stopped and the seashore were sufficiently deep. In that case we could neglect the effect of the current. But this cannot be done when the depth of the sea is not too great. At a level just above the bottom, equal to the friction depth D , the current changes by an absolute amount and direction similar to that which occurs in the case of a purely drift current. In this case friction, arising between the sea bottom and the levels of water just above the bottom, plays the part of the wind. The analogy between these phenomena becomes clearer if we represent as the moving element not the water relative to the sea bottom, but the sea bottom relative to the immovable water.

At the level just over the sea bottom the vector of current speed deviates 45 degrees to the right of the pressure gradient or to the left relative to the direction of the subsurface gradient current; the further from the bottom, the more the speed vector deviates to the right, increasing in absolute value. Finally at a distance D' from the bottom it deviates 90 degrees to the right of

the pressure gradient and has the maximum absolute value.

Furthermore, in measuring the increase in the distance from the bottom to the surface of the sea this value does not change. (Figure 4).

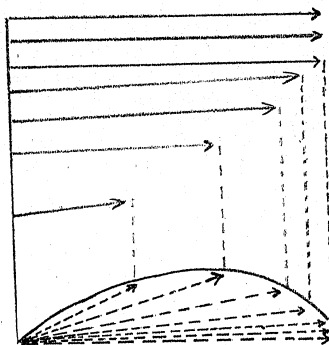


Figure 4. Diagram of changes in current speeds at the level D' of the bottom.

When the sea level is less than the friction depth and is only about $0.25D$, the absolute current speeds substantially change with the depth, but the direction of the current remains at all depths practically unchanged, insignificantly deviating to the right of the pressure gradient.

At a sea depth of about $0.50D$ the surface current vector de-

viates 45 degrees to the right of the pressure gradient. Changes in directions with depth, as in the first case, are not great, but the speeds change substantially with depth as before.

In nature, as a rule, drifts and gradient currents are observed simultaneously. In addition, only in special cases can the currents be looked on as steady. In view of the complexity of studying the unsteady motions it is best to examine first the diagram of steady currents.

If a sea with a depth of $H \geq 3D$ bordered by a shoreline is under influence of a uniformly steady wind, then it is unavoidable that a slope in the physical surface of the sea will develop. Under these conditions the entire water mass is divided into systems of active currents in three layers.

1. The upper layer of thickness D , in which the purely drift currents move as a result of wind action. If we assume that the wind speed is drawn along the Y axis, then the full flow of the water Φ of the purely drift current in layer D will be along the S axis, that is, it will be perpendicular to the wind direction. The discharge of water of this flow can be expressed by the relationship:

$$\Phi'_x = \frac{U_0 D}{\pi \sqrt{2}}$$

The discharge in the direction of the Y axis becomes an even zero:

$\Phi'_y = 0$. Consequently, as a result of wind action arising in layer D the flow is such that the general resultant flow of water in the

entire layer is moving in the direction perpendicular to the wind action.

2. The meeting of this flow with the move causes the surface^e_λ of the water to have a slope and develops a hydrostatic pressure gradient.

As was examined above, under the action of a pressure gradient in the entire water mass from the surface of the sea to the bottom of gradient flow is developed directly at right angles to the pressure gradient. This current has the same speed from the surface to the bottom. The full flow of a gradient current can be written as:

$$\Phi'' = \frac{g \gamma H}{2 \omega \sin \varphi}$$

Here γ is the tangent of the sea surface slope to the horizontal surface.

3. In the sea bottom layer equal to the lower friction depth, a gradient current similar to the purely drift current is active. The gradient current has the role of the surface current in this case. If we suppose that the pressure gradient is along the X axis, then the flow along the X axis can be expressed as

$$\Phi_x''' = \frac{g \gamma D'}{4 \pi \omega \sin \varphi}$$

and the current along the Y axis is equal to zero:

$$\Phi_y''' = 0.$$

To obtain a full flow active in each of the three layers, it is necessary to add to the gradient flow of water moving in all three layers, currents caused by wind action in the upper friction depth layer and by the friction against the bottom in the lower friction depth layer, that is;

$$\Phi_D = \Phi'' + \Phi'; \quad \Phi_{H-2D} = \Phi'';$$

$$\Phi_{D'} = \Phi'' + \Phi'''.$$

The current speeds of a full flow in each layer will be expressed as:

$$U_D = U'' + U'; \quad U_{H-2D} = U''; \quad U_{D'} = U'' + U''';$$

$$V_D = V'' + V'; \quad V_{H-2D} = V''; \quad V_{D'} = V'' + V'''.$$

In this equations U_D and V_D are the component summary speeds acting in the upper friction layer D; U'' and V'' are the component gradient current speeds; U_{H-2D} and V_{H-2D} are the component speeds active in the middle depth layer; $U_{D'}$ and $V_{D'}$ are the component speeds active only in the lower friction layer.

Above we examined some of the results of the theory of drift and gradient currents in the supposition that the flow is steady.

In unsteady purely drift currents a distinct time is required

in order that the currents studied on different horizons will appear close to being steady in absolute value and direction. The time required to study the current increases with depth. On Figure 5 are shown the successive positions of the ends of a speed vector for a surface current from the beginning of a wind action.

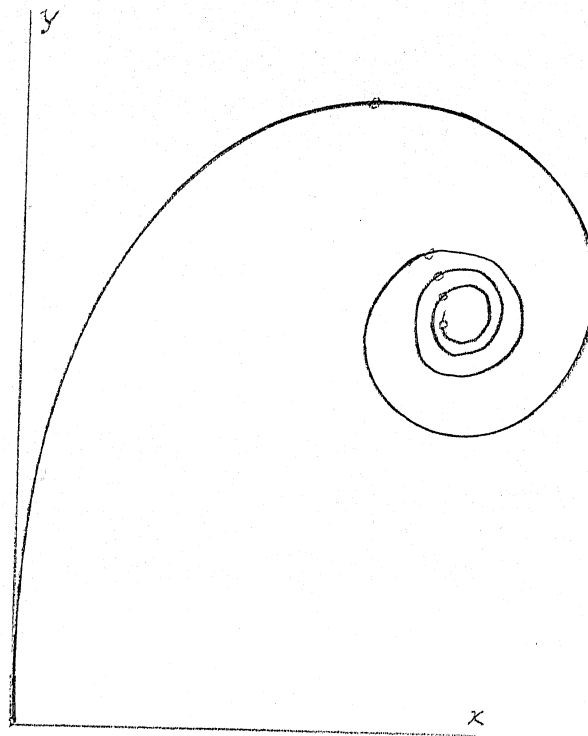


Figure 5. The change in the unsteady surface current vector. The points shown on the diagram indicate the successive positions of the end of the vector drawn from the origin of the coordinate through 3, 6, 9, etc. hours after the beginning of the wind action.

Unsteady gradient current also varies for a long while in absolute speed value and in direction around some mean speed value, corresponding to the steady gradient current vector.

11. EMPIRICAL STUDIES OF THE RELATIONSHIP OF CURRENTS TO THE WIND

The theory examined does not give directly a quantitative relationship between current speeds and the wind. This dependence must be obtained empirically. But empirical studies directed at clarifying this relationship are connected with great difficulties. First of all the difficulties arise out of the complexity of erecting observation points far from the shores at great depths for a long period of time. But prolonged observations are just what is necessary for this purpose. We cannot limit ourselves to episodic studies for the clarification of the entire picture of currents in various typical cases.

The necessity of studying a large number of observations leads us to the utilization of observations made by floating lighthouses (Often on floating lighthouses daily observations are made 2-6 times a day of the currents on the surface of the sea.) to establish the dependence of the current speed on wind speed. The sea depths where these observations are made are insufficient as the floating lighthouses are usually placed in shallow waters or are dispersed somewhat close to the shore. Taken all together they make a study of this dependence difficult even under the most favorable circumstances.

In accordance with the research of many authors the relation be-

tween the surface current speed U_0 and the wind speed has a simple form:

$$U_0 = \frac{A}{\sqrt{\sin \varphi}} V.$$

The mean value of the wind coefficient A resulting from the studies of several authors equals 0.0127, if U_0 and V_0 are measured in meters per second. Unfortunately, this value cannot be considered as sufficiently dependable for the following reasons. The working out of observations when establishing coefficients suffered in that either synthetic, purely formal methods of excluding residual currents were applied (Pal'men), or as a result of averaging a great number of observations it was supposed that the residual current speed becomes zero in view of the variations in directions of the residual current (N. N. Struyskiy, S. Ya. Shcherbak, V. A. Zenin).

V. A. Zenin notes that according to studies of the current observations on the Caspian Sea the relation between the current in knots and the wind (in meters per second) for the mean latitude of the Caspian is expressed as $U_0 = 0.04V$.

The conversion of knots to meters per second show that the wind coefficient in this case is equal to 0.02 and not 0.0127.

Yu. N. Neronov analyzed the products of observations somewhat differently. The method of research applied by this author has a definite advantage in comparison to the mechanical averaging which, unfortunately, is used too widely. Neronov made studies of the products of 12 years of observations of the currents on the floating lighthouse, Adlergrund, where the current and wind were observed

every 4 hours. This lighthouse is located 35 kilometers from the nearest island of Bornhohn on the Baltic Sea. The prevailing sea depth in this area is 20 meters. The geographical position of the lighthouse is such as to permit the supposition that the influence of the shores and the relief of the bottom for all wind directions is unchanged.

First the question of how long the period of time should be for a current to acquire the characteristic of steadiness was examined.

It would seem that the simplest way to decide this question would be to select from all the observations those times in which the wind suddenly arose after a prolonged calm (not less than 20 hours), and then acted for a prolonged time (around 20 hours) unchanged in force and direction. Then, using the products of observation after 4 hours we could follow through how the currents change with time, and in this manner determine from what time the currents, having reached a definite speed and direction, do not change further.

However, despite the large number of observations, such desirable cases were not discovered. Success was met in finding only a case in which the wind acted for a prolonged time (20 hours), substantially unchanged in speed and direction, independent of what the wind was like over the sea prior to these 20 hours. Consequently, in the indicated conditions towards the beginning of the 20 hour period of time some sort of ~~current caused by wind action~~ acting before the observation time was already being observed at sea.

But whatever wind there may have been until this time, when it was determined in force and direction, from the theoretical calcu-

lations one can assume that for 20 hours the action of the unchanging wind on the sea surface current also takes on the characteristic of being steady.

To verify this situation Neronov compared the current speed observed after the time of establishing the westerly and ~~S~~^{Southwest}erly wind with currents observed after 16, 12, 8, and 4 hours. It appeared that the relation of speeds in the 4 hour periods of time which divided the observation is rather close (Figure 6). In most of the cases the dispersal of dots is not over ± 7 centimeters per second, which does not exceed 20 percent of the amplitude of current speeds observed in this area.

In the 8 hour intervals that divide the observations, the final speed U_2 actually is distinguished from the initial U_1 . The dispersal of dots appears large (Figure 7), but such that the field of the graph can be split into sections by a line: filled and not filled by dots. Such a definite division of the field into two parts shows the limits (corresponding to the straight line marked 8 hours) beyond which the final current speeds U_2 cannot be observed corresponding to the values of the initial speeds U_1 .

Increasing the time intervals dividing U_1 and U_2 to 12 and 16 hours naturally makes the dots on the graph occupy more and more area (see the lines marked 12 hours and 16 hours on Figure 7). The position of the line marked 16 hours indicates that during a 16 hour interval of time dividing the initial and final speed, the final speed can change within the limits of from zero to 40 centimeters per second, that is, to the maximum speed value observed in the established winds. This circumstance allows: a deduction to be made:

16 hours is sufficient for the unchanging wind acting during this time to determine fully the speed of a drift current regardless of the starting speed.

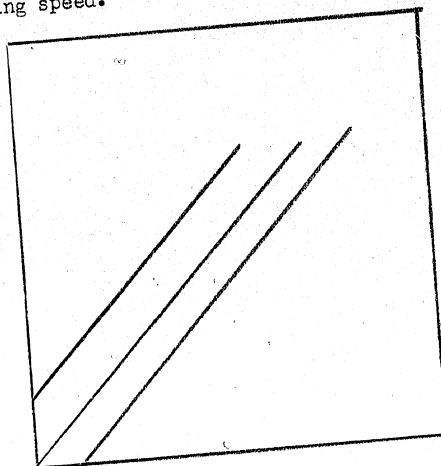


Figure 6. Relation of current speeds during a steady wind. U_2 is taken 4 hours after U_1

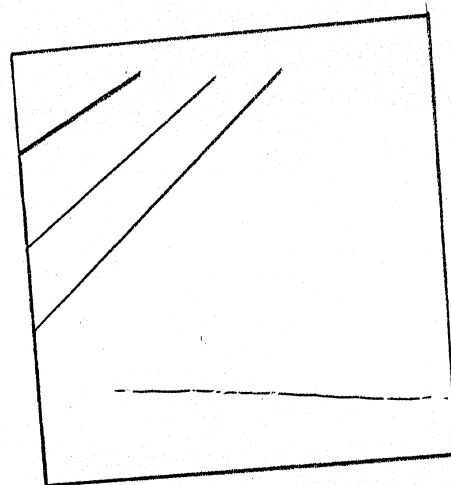


Figure 7. Relation of current speeds during a steady wind. U_1 and U_2 taken over 4, 12, and 16 hours.

Having determined the time necessary for the dying down of the initial current speed, or the minimum period of time after which the current may be considered steady, Neronov picked out of 10 years of observations 38 instances when the wind acting for 16-20 hours could, for practical purposes, be considered unchanging in force and direction. In this we cannot disregard the difficulty of finding a great many cases, because for a period of 16-20 hours the wind is seldom steady, especially if its speed is either insignificant or very great. Having used the observations picked out in this manner he constructed a graph relationship between the force of the wind in balls of Bofort and the current speed in centimeters per second (Figure 8).

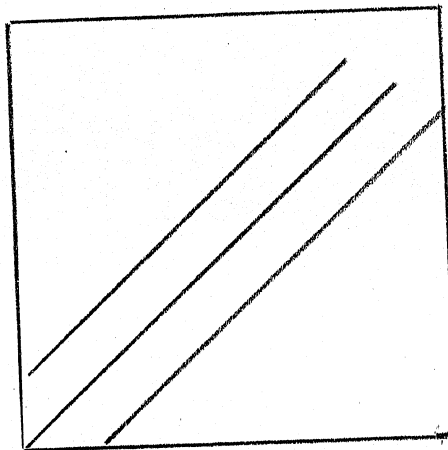


Figure 8. Relation of wind force (in balls) and current speed (centimeters/second).

The current speed calculated by this formula in 92 percent of all bases varies, not over 7 centimeters per second, from the speed

actually observed.

Transferring balls of the Bofort scale to meters per second (the transfer is made by the Simpson Table recommended by the international meteorological commission in 1926) causes the relation to appear nonlinear. This indicates the retarded growth of the current speed in comparison to the wind speed (Figure 9): $U = 0.04 V^{0.82}$.

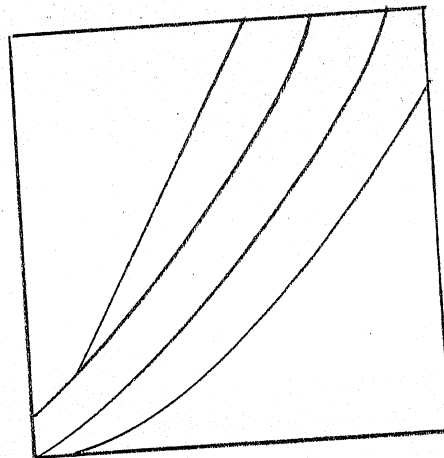


Figure 9. Relation of wind speed (in meters per second) and current speed (in meters per second). For comparison a line is shown on the figure equal to:

$$U = \frac{0.0127V}{V \sin \varphi}$$

The probability of not producing errors of ± 7 centimeters per second in the equation in this case is also 92 percent.

The construction of similar dependences for intervals of time when the steady wind acted less than 16 hours proved to have considerably less probability, specifically:

Time of steady wind action, hours	Error meters/second	Probability percent
16	0.07	92
12	0.07	86
8	0.07	80

When the force of a newly developed wind was compared with the current speed, then for the same error the probability was only 52 percent (Figure 10).

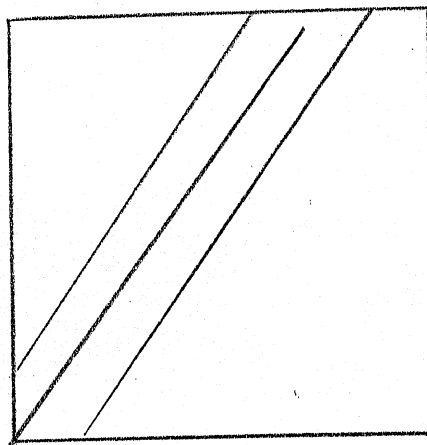


Figure 10. Synchronized relation of wind force and current speed.

Data on the probability of the relationships obtained in this manner confirm that the time interval of 16 hours determined by comparing initial and final speeds on correlated graphs, is indeed sufficient to consider a current steady.

The examined study of current dependence on the wind as observed from the lighthouse Adlergrund is really distinguished by its methods and results from those obtained earlier.

For example, having studied this dependence as well as the products of observations from the lighthouse Adlergrund, Dinklage and Neyman obtained accordingly the following values of wind coefficients: Dinklage - 0.0127, Neyman - 0.0204. Both of these researchers attempted to determine the wind coefficients by averaging observations. The former used observations for only one year, and the later for 9 years. Therefore, the results obtained by them cannot be close to reality. The method of averaging in this case cannot lead to positive results.

Calculations of current speed values by Neronov's relationship have approximately twice the significance as by a wind coefficient 0.0127 (see Figure 9).

In accordance with Ekman's theory, not examined here, the summary current speed observed on the surface of the sea essentially depends on the relief of the bottom and the configuration of the shores, the depth of the observation area, distance from the shore, etc. The empirical proof of this situation can be adduced, based on the observations being made on floating lighthouses.

Having analyzed the products of observations of currents from a lighthouse under conditions unsuitable for developing drift currents, similar to the observations from the lighthouse Adlergrund, a relationship was obtained between the current speed and the wind, represented by a wind coefficient very close to 0.0127.

The graph representing this coefficient (Figure 11) shows a weaker relation between the wind and the current than on the lighthouse Adlergrund both in the dispersal of dots and in the slope of

the line. Using, however, the formula $U = 0.04V^{0.82}$ in calculating the currents observed from the lighthouses Ovizi and Nekomangrund, located in conditions similar to those of the lighthouse Adlergrund, indicated its full adaptability to current characteristics in this area. All this points to the necessity of establishing empirical dependences calculating the local conditions, that is, to construct, practically, these dependences separately for each area with characteristic peculiarities of bottom and shores.

No fewer difficulties arise in studying under natural circumstances the relationship of the current direction with the wind speed and direction. A basic reason for the difficulties are the very limited opportunities of making prolonged observations in deep areas of the sea. Most of the work devoted to this problem indicates that the deviation of the drift current from the wind is about 2 points of the compass, that is, about 22 degrees. However, such a conclusion cannot be considered as inconsistent with theory, since the observations made as a rule in shallow areas were thoroughly analyzed.

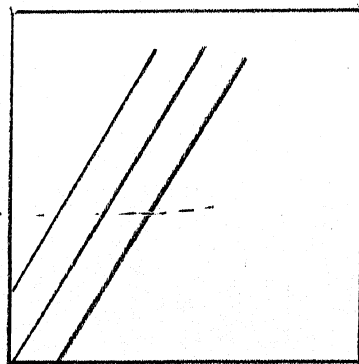


Figure 11. An example of the relationship of the wind force and current speed in circumstances unfavorable to the development of drift currents.

The statistical compilation of 6000 measurements of currents and wind on the Caspian Sea can be reduced to the following data, derived from the work of V. A. Zenin:

Wind (balls)	2	3	4	5	6	7	8	9
Deviation of current, (percent)	67	63	52	42	37	34	20	5

Here we have the percentage of current deviation, caused by the wind, from the wind direction of more than two points of the compass. From the given data it follows that with an increase in wind force the percentage of deviation becomes less and less. Such a characteristic decrease in percentage with an increase in wind force agrees with theory. With an increase in wind force, the current speed also increases, as does the number of water eddies, represented by the coefficient μ , which in turn is related to the friction depth D so that the less the relation of the general depth of a sea to the friction depth D , the less is the angle of current deviation from the wind direction. Consequently, the sea appears to grow shallow with an increase in wind speed, and therefore it is natural that even the percentage of current deviation from the wind direction becomes less.

What has been said about empirical and theoretical studies of the dependences of current speeds and direction on the depth and wind speed confirms that before beginning forecasts of drift currents in the entire or in one of the areas, we must work considerably to establish the dependences in which the basic physiogeographical peculiarities of the area would be considered. Expanding the dependences for similar conditions can cause considerable errors.

To establish dependences, similar to those examined above, a large number of good-quality observations are necessary which at present are available in a limited number and the creation of special observations for this purpose requires great resources and effort. The theoretical solution to the problem of the dependence of drift currents on the wind for the real physiogeographical conditions is also still meeting with unsurmountable difficulties.

Therefore, an attempt was made to avoid these difficulties and to give perhaps a very approximate, but a general solution to the problem of the dependence of currents on the wind action with the help of the products of episodic observations of currents in one of the seas. For this purpose one resorted to establishing the typical conditions of wind currents over the sea. Several types in which the most frequently observed actual wind currents and the synoptic situations related thereto are to some degree of accuracy included, were established.

Then it remained to be clarified, which types of currents in the entire sea are related to definite types of synoptic conditions. This problem was resolved on with the aid of all the observations of currents which were ever made. Such observations, unfortunately, proved to be too few -- altogether around 10,000, which becomes apparent, if we recall, that in solving this problem for just one point -- lighthouse Adlergrund -- products of 12 years were used, where each year produced no less than 1500 observations. Still, a bold attempt at a solution to this problem was undertaken. All the observations of currents relative to this or that area were, in relation to the synoptic conditions, classified into types and placed on charts. Further, on the basis of the conclusions of

the drift current theory white spots were filled in and diagrams of currents were constructed corresponding to the flows of wind that had been determined.

There is no doubt that the diagrams of currents made in this manner suffer from very substantial defects as does the method of typing, widely applicable and applied in synoptic meteorology but not giving any amount of objective and dependable results. Nevertheless, such an attempt shows a commendable interest and is a step forward in comparing rose ^ecurrents with the conventionally constructed charts.

With the aid of these diagrams of currents and forecasts of the barometric and wind conditions over a sea, attempts were made at forecasting currents; however, thanks to the absence of observations of currents at sea, no success was achieved in getting an objective characteristic of justification of this forecast.

It stands to reason that this method of forecasting as any other method not tested on any large amount of empirical work, cannot be recommended for adoption in operational work. It is necessary to make every effort to complete it and particularly to make it objective and to express quantitatively the peculiarities of atmospheric processes as well as the currents related thereto, at least for the most important sectors of the shipping lanes. For this the most effective means has not yet been utilized -- relative to tide and currents. In the section on the drift and gradient currents theory we mentioned the very close interdependence between the slope of the sea surface and the current elements. Not having the necessary observations of currents, we must attempt to supplement their short-

comings with a joint analysis of ^ssea level variations observed
from shore stations and posts, with the existing or special,
even though fragmentary, observations of currents for various wind
and barometric conditions. Only after completing such an analysis
can one expect navigators to make practical operational use of the
forecasts of currents for large portions of a ship's course.

12. A STUDY OF NONPERIODIC VARIATIONS IN SEA LEVEL

The drift and gradient currents theory allows us to analyze the interrelationship between wind action, currents and tide variations occurring in deep waters. If a sea has a depth of not less than 3D and steep and deep shores the theory allows us to determine quantitatively the relationship between variations in sea level and currents. Qualitatively the relationship between these elements of the pattern are as follows: if the wind is blowing perpendicular to the shoreline from the sea inland or from the shore seaward then the level at the shore does not change in this case, because the current arising from the wind action moves in a direction perpendicular to the wind direction, that is, parallel to the shore.

Conversely, when the wind is blowing along the shore from left to right or right to left relative to an observer facing seaward we have either a rise or fall in the level as the case may be. If the wind is blowing from left to right then the flow of water caused by the wind action will tend shoreward and as a result there will be observed a running-up on the shore -- the sea level will rise. In the case of the wind being from right to left the current will be directed from the shore seaward and consequently the level at the shore will be lowered, that is, a running-out phenomenon will be observed. When the wind acts at some angle to the shoreline then the amount of running-out or running-up will be determined by the projection of the wind on the shoreline.

An entirely different picture will be observed if a steady wind, whose force and direction remain unchanged over the entire surface of the sea, will act over the surface of a shallow closed-in sea (Figure 12).

In this case, due to the viscosity, the upper layers of water will be attracted in the direction of the wind. At the shore onto which the wind is blowing the waters carried by the running-up flow will be accumulating. At the same time this current will decrease the amount of water at the opposite shore. As a result, at the windward shore there will be a rise in the sea level, but at the lee shore there will be a fall. A pressure gradient will arise directed against the wind. Since it is assumed that the wind over the entire surface of the sea has an equal force and direction, a current directed leeward cannot arise. This current arises in the subsurface layer and is directionally opposite to the surface current and to the wind. Naturally, the stronger wind brings the surface current to high speeds, resulting in a greater slope on the surface of the sea, that is, a greater pressure gradient, which in turn will cause great speeds in the subsurface water layers directed to the side opposite the wind.

It is quite obvious that between the wind speed, the pressure gradient, the subsurface and surface currents, equilibrium must be established under which there will be a full compensation of the running-up of the waters by the subsurface current. Such a calculation establishing the circulation of waters under wind action^{was} made by V. B. Shtokman for sea areas distant from the shore.

V. B. Shtokman established what relationship the friction force, arising as a result of wind action on the water surface, T , the depth of the sea H , the water density ρ , the slope angle of the sea surface to the horizontal, caused as a result of wind action, γ ; must have^{to} one another. This relationship has the form:

$$\tan \gamma = \frac{3T}{2g\rho H}$$

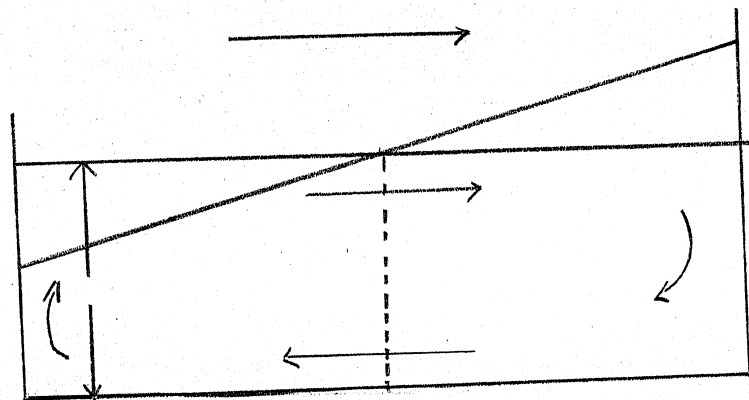


Figure 12. Diagram of the development of sgonno-nagonny phenomena in a shallow sea.

The current speed related to the slope angle of a sea surface is shown by the following relationship:

$$u = \frac{g\rho \tan \gamma}{2\mu} (H^2 - Z^2) + \frac{T}{\mu} (H - Z)$$

In this and the preceding relationships g is the acceleration of gravity; μ is the coefficient of turbulence friction; Z is the vertical coordinate directed downward (Figure 13). The surface current speed in this case is expressed rather simply:

$$u_0 = \frac{\tau H}{4\mu}$$

It is directly proportional to the acting force and the sea depth and inversely proportional to the coefficient of internal turbulence friction.

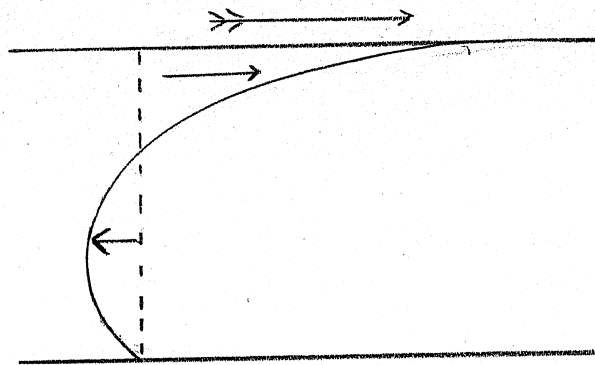


Figure 13. Curve of current speeds for a steady wind in a shallow sea.

It is not difficult to establish that the border between two directly opposite currents (surface and subsurface) comes to $1/3 H$ from the surface. The upper layer has the running-up current, and in the lower layer, consisting of $2/3 H$, the compensating current acts in an opposite direction. The curve of the speeds is shown in Figure 13.

The theory developed by V. B. Shtokman, unfortunately, cannot be applied to coastal areas of a sea. It is true only for shallow sea areas distant from the shoreline. A further development of the theory of sgonno-nagonnye variations in sea level was achieved

in the works of L. S. Leybenzon and N. A. Bagrov. The theory evolved by N. A. Bagrov allows us to follow the development and the subsiding of the wind running-up phenomenon relative to an unexpected development or dispersal of the wind. How the level behaves in the case of a sudden wind development is shown on Figure 14 by curve a, while the case of a sudden wind dissipation is covered by curve b on the same figure. On this figure we have the relationship ζ/ζ_0 , in which ζ_0 , represents the level slope from a position of equilibrium in the case of a steady wind speed. This value can be calculated easily with the aid of the relationship:

$$\tan \gamma = \frac{3T}{2gPH},$$

Since

$$\tan \gamma = -\frac{z}{x}.$$

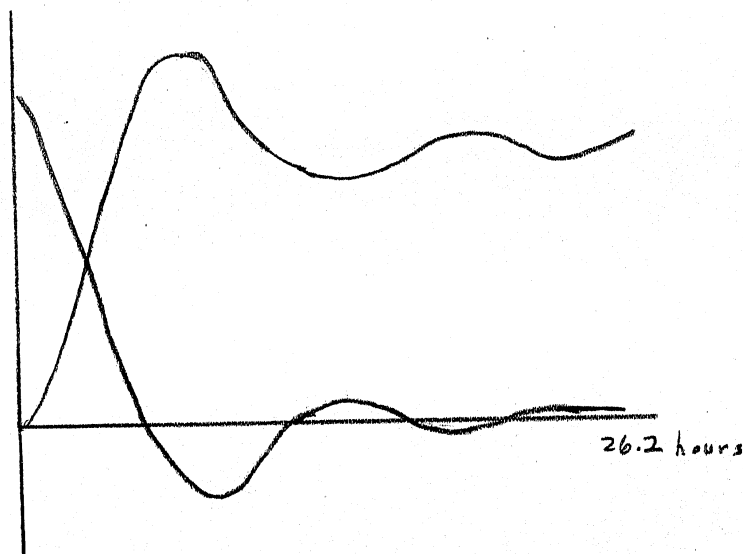


Figure 14. The change in sea level during the time of a sudden development (a) and a sudden dissipation in the wind (b).

The letter ζ designates level slope from a position of equilibrium for an unsteady movement. This figure shows that in case of an unexpected wind development in the direction of the running-up phenomenon the level rises sharply to a value somewhat higher than that which would be observed in the case of a steady wind flow of the same speed, then drops somewhat lower than the position of a level under stationary conditions.

The level variations occur near the value which it would have if the wind acted in the same direction with unchanging force for a prolonged time. These variations in time become less and less in amplitude until finally they subside completely: These variations in sea level somewhat close to the mean value for a stationary condition can be considered as a seiche, that is, as free standing waters. Figure 14 is derived from the work of Bagrov, constructed to be applicable to the conditions of the Sea of Azov without the Bay of Taganrog. The length of the water is taken as equal to 2×10^5 meters, and the depth 10 meters. For these values of length and depth of sea the seiche period caused by a sudden development or dispersal of the wind is equal to 13.1 hours. For this time the amplitude of variations occurring with this period diminishes to such a degree that it becomes not over 10 percent of the original value. Doubling the length of the basin also doubles the period of variations, but the time required for the dying down does not change with an increase in the period, that is, despite the increase in the period over the same 13.1 hours, the amplitude of variations will not be over 10 percent of the original amplitude.

Figure 15 shows the changes in current speeds along the vertical, relative to the dying out of the running-up phenomena when the wind stops, and Figure 16 shows the development of the

running-up currents at sea when the wind suddenly appears. Around each speed curve shown on these figures, the numbers indicate the time in hours counted in the first case from the time of a sudden wind dispersal, and in the second from the time of a sudden wind development. The positive values of the abscissa on Figure 15 and 16 are the speeds of water movement coinciding with the wind direction, and the negative ones are the opposite direction. In the beginning of the running-up phenomenon (Figure 16) only the upper layers of the sea move in the direction of the wind. In one hour the speed is noticeable already at all layers and only after about 5.2 hours do the subsurface countercurrents arise. It is extremely important to note that the speed of the surface current in one hour after the beginning of the wind action already attains the speed value of a steady current. Almost after 3 hours the speed of the surface current is over twice greater than the speed of a steady current remaining around this value with practically no variations.

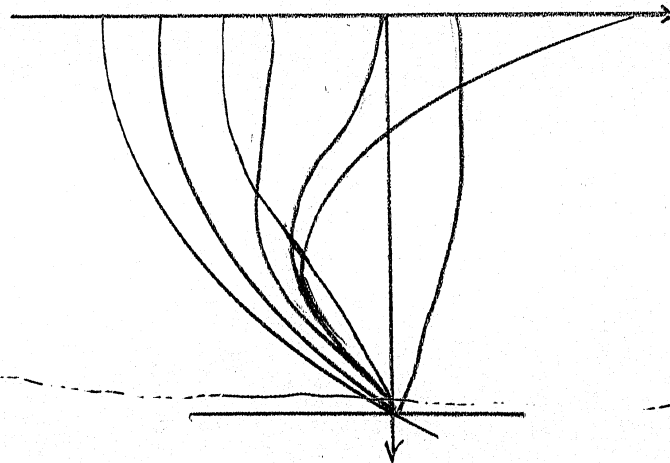


Figure 15. Curves representing the changes in current speeds after a sudden wind dispersal. The values next to the isoline indicate the hours passing from the time of wind stoppage.

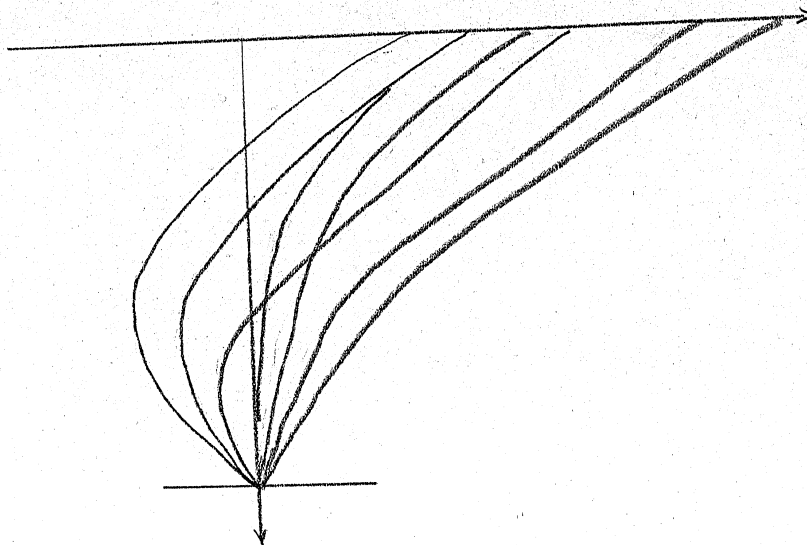


Figure 16. Curves, representing the changes in current speeds after a sudden wind development. The values next to the isoline indicate the hours passing from the time of wind development.

The dying out of wave variations, according to Bagrov, is approximately proportional to the square of the depth. The speeds at which the wave movements die out are so great that we can practically always disregard the initial speed distribution and the initial level.

In the examined observations of Shtokman and Bagrov the Coriolis force was not taken into account. It was assumed that its role in comparison with the friction force is so small that it can be ignored.

P. A. Kitkin, solving the same problem for a round sea with a parabolic bottom, considered the turbulence friction and Coriolis

force. His solution like Bagrov's solution, indicates the development of free variations -- seiche -- resulting from the influence of the wind on the sea surface. Under the influence of the Coriolis force there is a clockwise rotation in the plane of the free variations in the level, but the direction in which we note a rise in the level, deviates from the direction of the "running-up" wind. The time between the beginning of the highest level and the maximum wind force is not constant and depends on the wind acceleration. The amplitude of free variation according to Kitkin, as according to Bagrov, dies down though considerably less so.

Kitkin gives several examples of variations in the level of the Aral Sea in support of his theory. The conclusions made by Kitkin paint a considerably ^y more complicated picture of sgonno-nagonnye phenomena in comparison to the theoretical studies of other authors and in comparison to the representation made as a result of the empirical studies. Therefore, his theoretical position must be confirmed in large measure not by selective, as he does, but by continuous observations for an entire season.

In particular there is doubt that in shallow seas, like the Aral, Azov and Caspian (northern part), it is possible to have considerable deviation in the running-up phenomenon from the wind direction or a prolonged existence of free variations in the current of several periods simultaneously with the rotation of the plane in which these variations take place.

If such a phenomenon were observed in nature then the deviation of simple linear relationships between a gradient pressure and nonperiodic variations in the level which will be described in detail

below, would be impossible. Besides this, the forced variations caused by a periodic change in breeze winds would not give such a simple and correct form (Figure 17). The rotation of the plane of variations more complicated.

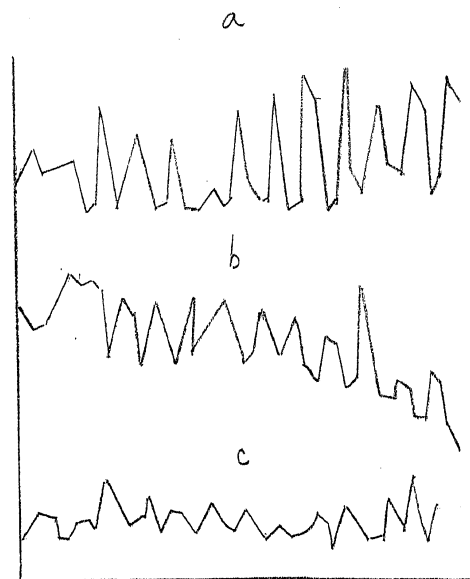


Figure 17. Variations in the level of the Sea of Azov maintained by breezes, June 1940.

a - difference between the air and water temperatures, b - the level in Taganrog, v - the level in Genichesk.

G. S. Ivanov compared the correct daily variations in the level of the Sea of Azov with the difference in air and water temperatures. The good coincident path of these curves served as a basis for him for confirming that these variations are compelled and the breeze winds are the compelling force. Usually, such variations in the level of the Sea of Azov are observed during cloudless skies and small barometric gradients. This phenomenon occurs in other areas too. N. A. Berlinskiy studied the correct shift of currents at the mouth of the Dnepr during cloudless weather and small pressure gradients. The reason for the shift in currents, apparently, is hidden in the same correct shift in breeze winds. Probably, this phenomenon is no less distinct in the Aral Sea than in the Azov, inasmuch as cloudless weather with weak current gradients is not rare. It is highly possible that this phenomenon led Kitkin to his misconception. He took the forced variations maintained by periodic breeze winds to be free variations. Figure 18, derived from his work, resembles Figure 17, taken from Ivanov's work. The period of variations shown in these figures is equal to the period of breeze winds. This example shows how it is sometimes difficult to determine the real reasons for phenomena observed at sea. [Figure 18 shown on following page]

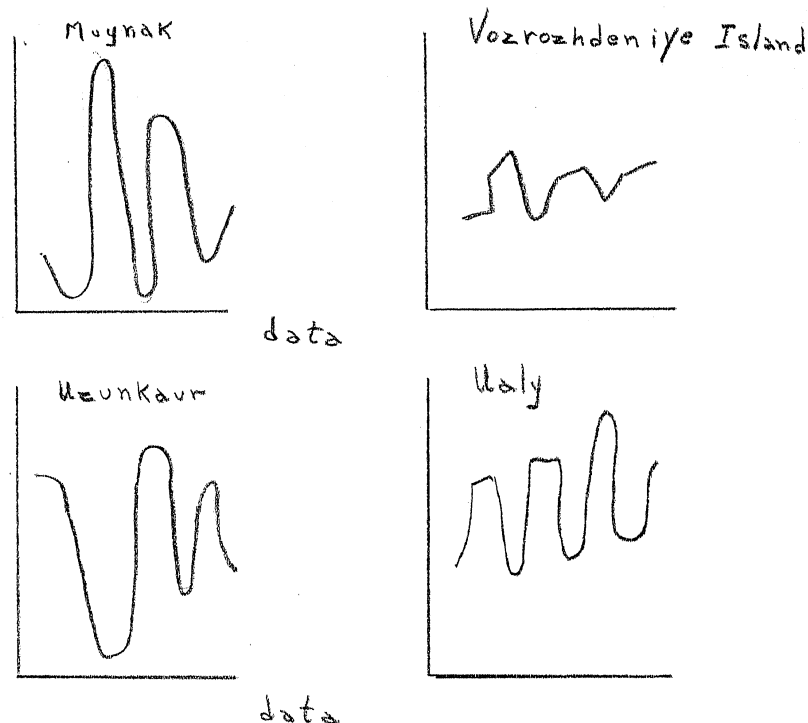


Figure 18. Variations in the level of the Aral Sea, recorded simultaneously by four stations

13. A SYNTHESIS OF EMPIRICAL RELATIONSHIPS FOR FORECASTING NONPERIODIC VARIATIONS IN SEA LEVEL

As noted above, predicting sgonno-nagonnye variations in sea level has great practical significance for shallow areas of a sea. ~~As a result in most empirical studies of sgonno-nagonnye~~ variations shallow areas are examined. Many observations were begun by clarifying the role of local winds in forming running-out and running-up phenomena at this or that point. Consequently it was concluded that the local wind has little to do with variations in sea level. The wind pulsation, the influence of wind speed and direction of local conditions and the lack of relation between the

wind observed at a separate point and its action over most of the water surface make such a conclusion entirely natural.

Above we examined the most simple cases when the wind does not change either in time or expanse, however, in nature, as a rule, the wind changes in time as well as in sea areas. Complicating the problem somewhat, one can imagine a case when the wind acts over the sea surface with a steady speed and direction, while the wind speeds differ. At one shore of the sea the wind is weak and at another it is stronger, even though the wind direction is alike everywhere (Figure 19)

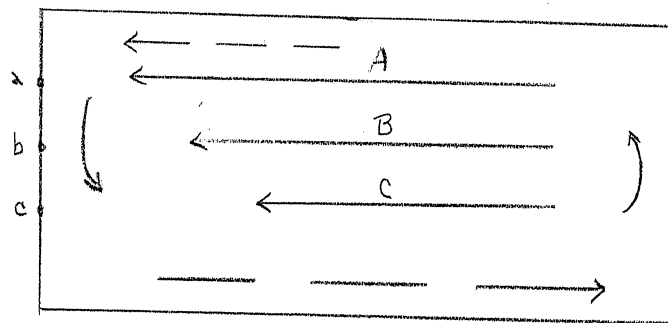


Figure 19. Diagram of currents in a closed shallow sea for a cross and uneven wind.

Shtokman, theoretically examining the question which currents would arise from such a wind, showed that in this case a current directed against the wind arises on the surface of the sea. And actually, this phenomenon is observed quite frequently in nature. In the indicated cases it is most difficult to determine the relation between wind action and current. But is it correct to consider the transverse wind inequality if the establishment of a relation not

between the wind action and the current, but between the wind action and the level variations is the problem of the investigation.

Inasmuch as it is difficult to solve this problem theoretically, it is necessary to use the products of observations and to solve the problem in rough approximation, but still with accuracy sufficient for practical purposes. At first we must assume that the transverse inequality of a wind playing such an essential part in the forming of countercurrents is practically nonexistent in determining the magnitude of level deviation from a condition of equilibrium at the shores. Figure 19 shows a rectangular basin over which a wind of one direction, but varying force, is acting. The wind force in a certain scale is shown by vectors A, B, C. We can assume that the shore depth gages a, b, c will fix the deviation of the level from a position of equilibrium, are in practice close to each other and in any case of the same sign, and proportional to the mean wind force B.

Determining the mean wind speed by indirect observations at shore points and in the open sea is difficult. To avoid this it is best to coordinate the wind force and direction with the atmospheric pressure gradient. Whereupon, if the pressure gradient is calculated at two maritime stations, then, actually, this gradient will resemble the average wind speed over a sea. Such a method completely removes local peculiarities in the wind system relative to the local relief or the shore configuration. However, relative to theoretical studies an unsteady wind system causes seiches to be developed which, even though they die out rather quickly, can attain a considerable value.

Atmospheric pressure also pulsates within definite limits. That is why in setting up the relation it is necessary to come to average in time for variations in sea level and for pressure differ-

ences. Averaging allows us to represent level variations and pressure changes as being slow changing, and this in turn allows us to examine all phenomena as being steady.

Hence averaging provides an opportunity to simplify the problem. In solving the problem of the time interval in which we must average level values and pressure differences, it is necessary to recall the results of Bagrov's theoretical studies, according to which water movement in practice comes close to being steady after 13 hours. Atmospheric pressure, as is well known, has a considerable daily movement. Therefore, averaging the pressure average should not be taken for a time interval of less than one day. A one-day period of time is also entirely sufficient for averaging the level.

Using the same studies of Bagrov, we can consider that any stable change in wind speed and direction will reflect on the change in sea level in a period of time not exceeding 13 hours. Empirical studies of this problem were made in the Marine Section of The Central Forecast Institute by S. I. Kan from the products of observations made on the Caspian Sea. The conclusions made by Kan agrees with the results of Bagrov's studies.

~~In this manner changes in sea level occur almost simultaneously with wind changes.~~ Therefore, it is impossible to formulate a forecast procedure for sgonno-nagonnnye variations in level without forecasts of the wind or air pressure.

The impossibility of formulating a forecast procedure for sgonno-nagonnnye variations in level without a forecast of wind or air pressure derives also from the results obtained in his time by

T. P. Maryutin. In developing a forecast procedure for nonperiodic variations in sea level for the mouth of the Severnaya Dvina T. P. Maryutin obtained several equations having a satisfactory reliability for advance periods of 6 and 12 hours. Increasing this advance period brought a considerable decrease in the reliability. In Maryutin's equation the pressure differences observed at the stations Tsip-Navolok and Kanin-Nos, and the level observed at the point for which the forecast is made are used as independent variables. Formulating the equation for forecasting even for short periods in advance (6 to 12 hours) became possible only as a result of an artificial method. All the observations being developed were divided into two groups according to the movement of the level: (1) the ascending level stage and (2) descending level stage. For each of these stages equations were formulated separately. Still, it is clear that such a division is easy to make only for the products of recent observations. Where mere equations are used for forecasts the solution of the problem of which of the equations, that for the drop or that for the rise, should be used, appears difficult. And in connection with this, the necessity arose of formulating forecasts of sgonno-nagonnye variations in level, relying on air pressure forecasts.

The formulation of a level forecast on a pressure forecast can be useful in practice only when the pressure forecasts will have a high reliability, and the relation between the pressure gradient and the level will be sufficiently close.

Practice has shown that a pressure forecast for a period greater than 12 hours made by the usual synoptic method has an insufficient reliability to solve this problem, whereas for 12-hour

periods in advance the forecast reliability increases considerably. Therefore, in the present status of the problem we cannot use the air pressure forecasts for level forecasts if they had an advance report period of over 12 hours. However, the circumstance that the level assumes a value related to the steady wind, in a period of 6-12 hours, in case of slow changes in pressure gradients, can serve as a basis for increasing the advance report period for level forecasts from 12 to 24 hours.

Besides this, an important part in improving forecasts of sgonno-nagonnye variations in sea level can be played by the fact that atmospheric processes have a definite stability in time.

As is known from Mul'tanovskiy's work and his school (T. A. Duletov, S. T. Pagav, A. A. Rozhdestvenskiy, N. E. Shirkin. The basis for a synoptic method of long range weather forecasting. Gidrometeoizdat, L. -M., 1940.), and from other studies (N. A. Belinskiy's Experiment in determining the index of atmospheric circulation. Works of NITs GUGMS, series V, issue 14, M. -L., 1946.) a sharp change in the sign of a barometric field occurs after about a week. Undoubtedly, this peculiarity cannot be reflected in the path of the level variations. The empirical studies of level variations confirm this condition. So, S. I. Kan compared with one another the sliding mean decade levels calculated from the observations at the cities of Baku and Kaspiysk (Figure 20). As the figure shows, a linear relationship between the level variations at these two points does not exist. This is so because in Baku where sgonno-nagonnye phenomena are insignificant, after getting the 10-day average the curve for level changes in time became very

close to a straight line fixing a monotonous level development connected with the seasonal path. In Kaspiysk, however, where the sgonno-nagonnye phenomena are outlined considerably more clearly on the curve of the path of mean 10-day level values parallel to the monotonous level growth, we note wavelike variations.

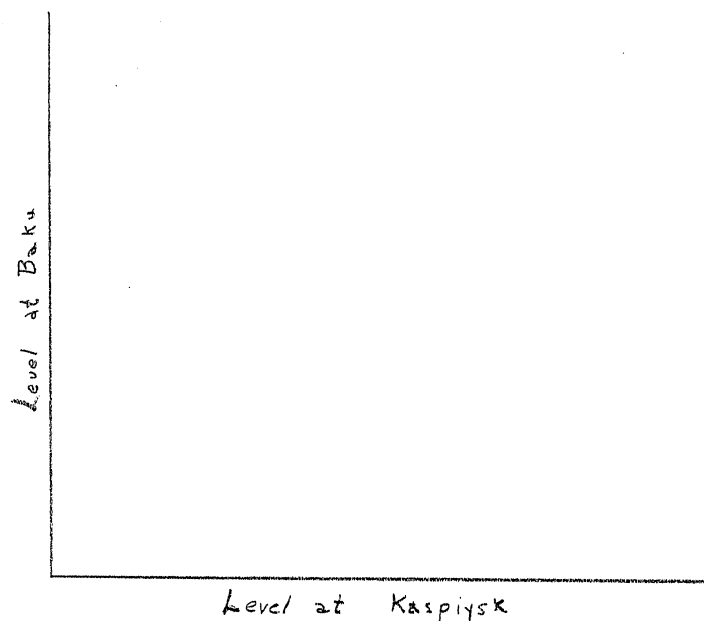


Figure 20. Relationship between the sliding mean 10-day sea levels at Baku and at Kaspiysk

Having calculated the mean 10-day values for pressure differences at the Gur'ev - Makhachkala stations, Kan succeeded in determining the synchronous relationship between the mean 10-day level at Kaspiysk and the pressure differences at Gur'ev - Makhachkala. Furthermore, with the aid of this relationship the level variations at Kaspiysk dependent on the wind action and excluded from the mean 10-day level values at this point were calculated. In this manner

she succeeded in excluding from the path of the level at Kaspiysk the wavy variations caused by the wind action. After this the curve for the path of the level at Kaspiysk had the same monotonous path as the curve for the path of the level at Baku, and this made it possible to construct the linear relationship between the path of the level at Baku and Kaspiysk, shown on Figure 21. With this it was possible to determine that in the seasonal path the level at Baku is about 15 days behind in relation to the path of the level at Kaspiysk.

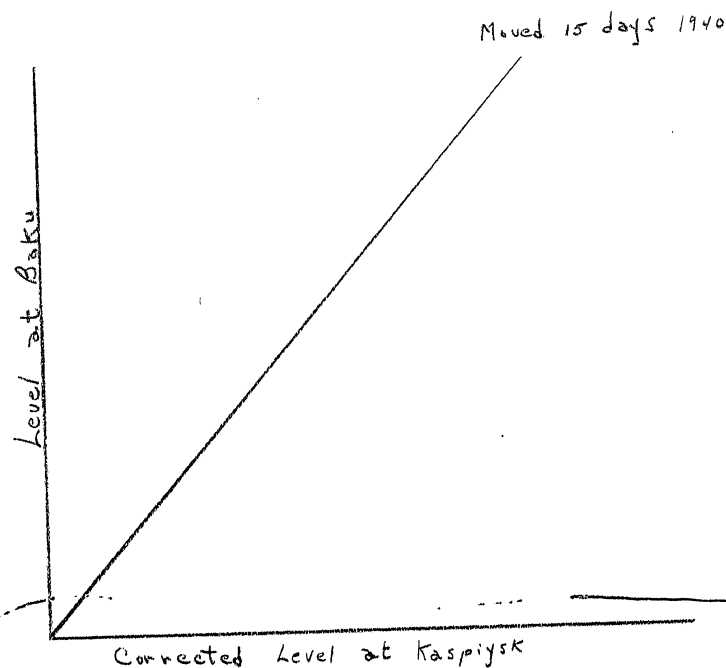


Figure 21. Relationship between the sliding mean 10-day sea levels at Baku and Kaspiysk after excluding the sgonno-nagonnye variations.

The examined relationships show that in the path of the atmospheric pressure and in the path of the level, together with the

comparatively short term changes, the variations of a large period are also noted. Generalizing this condition, we must remember that at each point of the seashore to some degree or other the annual path of the sea level dependent on the annual path of the wind over the sea must be expressed.

To determine the relationship, between the variations in sea level and the pressure gradients, the correct selection of the effective direction of a pressure gradient is most essential. Inasmuch as the height of a level rise depends not only on the force but also on the direction of the wind, it is necessary to select the direction of the projection of a pressure gradient relative to the wind direction for which the maximum "running-outs" and "running-ups" are noted. The selection of such a direction must be made separately for each point of level observations. In solving this problem two methods may be taken: that of matching and that of calculation. Kan, searching for the effective direction of a pressure gradient, used the matching method and had quite a laborious task. In determining the effective direction of a pressure gradient for Kaspiysk and for Peschanyy Island she had to calculate 10 coefficients of correlation. On Figure 22 the directions obtained by her are shown and the coefficients of correlation relative to these directions. The thick lines show the accepted directions for each of the named points. [See Figure 22 on following page]

This same problem could have been solved differently. We can always assume, on the basis of a theoretical concept, exactly which direction is a sgonno-nagonnye one. For an objective and a sufficiently accurate determination of this direction it is expedient to

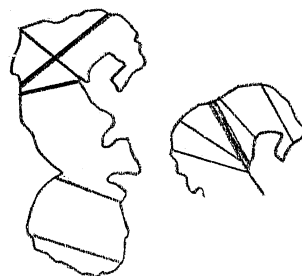


Figure 22. The directions of the pressure gradients and the coefficients of correlation related to these directions obtained in determining the relationship between the changes in the pressure differences and the level variations (A) at the Kaspiysk station and (B) at Peschanyy Island station.

calculate the pressure differences between two pairs of points located on two mutually perpendicular lines. Having named the pressure difference calculated between the first p_1 and the second pair of points p_2 , we determine two coefficients of correlation between the level variations and the pressure differences Δp_1 and Δp_2 . As a result we get the coefficients of correlation r_1 and r_2 . After which through the relationship $\tan \gamma = \frac{r_1}{r_2}$ we get a certain angle $\alpha = 90 - \gamma$, into which both selected lines must be set so that one of the coefficients will increase to

the largest value and the second decrease to zero. Actually it is very convenient if the direction obtained in this manner coincides with the line on the ends of which the marine stations are located. Further calculations are reduced to computing the pressure differences at these stations. The pressure differences are not divided into the distance between the stations, as is necessary when computing the pressure gradient, as this distance is constant.

When the direction of the effective gradient does not coincide with the stations we must determine two arbitrarily directed lines on the ends of which the stations would be located and determine the pressure gradient in the required direction by constructing vector diagrams. Using the vector method we can limit ourselves to three stations to calculate the projection of the pressure gradient on the effective direction for a given point.

If these preliminary calculations are made then we must next seek the relationship between the gradient (or simply the pressure difference* calculated for the effective direction and the level variations in the area being studied, in the form:

$$H = \bar{H} + k_1 (H_{\text{stand. obs}} - \bar{H}) + k_2 (\Delta \Gamma_{\text{forec.}} - \overline{\Delta \Gamma})$$

where H is the predicted mean daily level for a 24-hour period beginning 6 hours after the reading $H_{\text{stand. obs}}$ is taken; $H_{\text{stand. obs.}}$ is the level taken from the depth gauge at the time closest to the time of making the forecast; $\bar{H}$ is the sliding mean decade level; $\overline{\Delta \Gamma}$ is the sliding mean 10-day pressure gradient or simply the pressure difference between two points averaged for a 10-day period;

$\Delta \Gamma$ forecast is the pressure gradient or simply the pressure difference between two points computed in the pressure forecast for 12 hours in advance taken from the time for which $H_{\text{stand. obs.}}$ is taken; k_1 and k_2 are the coefficients determined for each point.

It is not difficult to see that a forecast with the aid of this equation is made on the calculation with a definite weight: tendencies in the path of the level $k_1 (H_{\text{stand. obs.}} - \bar{H})$ and the tendencies in the path of the pressure $k_2 (\Delta \Gamma_{\text{forecast}} - \bar{\Delta \Gamma})$.

If we assume that $H_{\text{stand. obs.}} - \bar{H} = 0$ and $\Delta \Gamma_{\text{forecast}} - \bar{\Delta \Gamma} = 0$, then we get $H = \bar{H}$. This one very important peculiarity allows us to use the equation not knowing the absolute mark on the depth gauge. In order to use these equations it is enough to know the mean daily levels for 10 days before making a forecast. This peculiarity of the equation is very useful in practice.

As was indicated above, in the path of the level and pressure we have a variation point with comparatively long periods and large amplitudes. Averaging for 10 days allows us to get a sufficiently close relationship between the path of a level and pressure which takes into account these variations. The inclusion of $\Delta \Gamma$ in the equation plays an essential part as it brings about a considerable decrease in the number of errors connected with the necessity of using pressure forecasts and the fact that the variations of a short period in the path of the level and pressure are more weakly related to one another than the variations of a long period.

Experience has shown the possibility of applying this method of forecasting in the most varied circumstances. For example, S. I. Kan determined this type of relationship for the area of the Volgo-

Caspian canal, the area of the Peschany Islands on the Caspian Sea. The actual and predicted variations in the level are shown on Figure 23, and the equation reliability is presented on Figure 24.

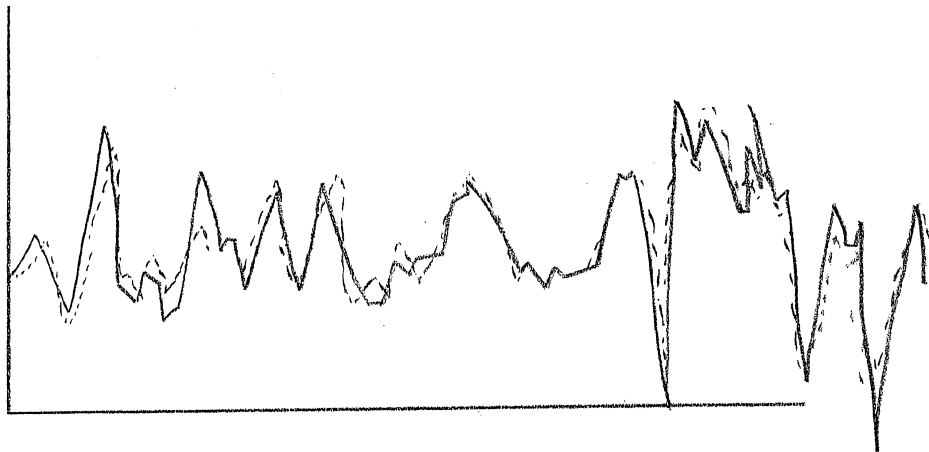
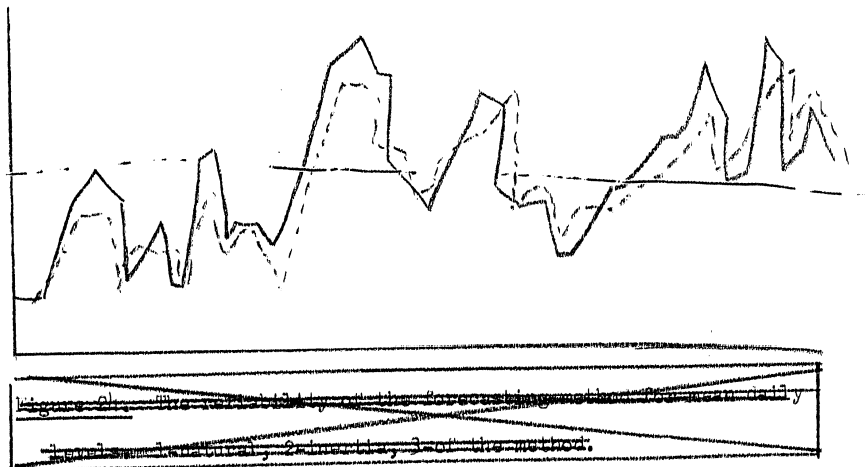


Figure 23. Actual (1) and precalculated (2) mean daily level at the Kaspiysk station (top) and at the Peschany Island (Bottom)



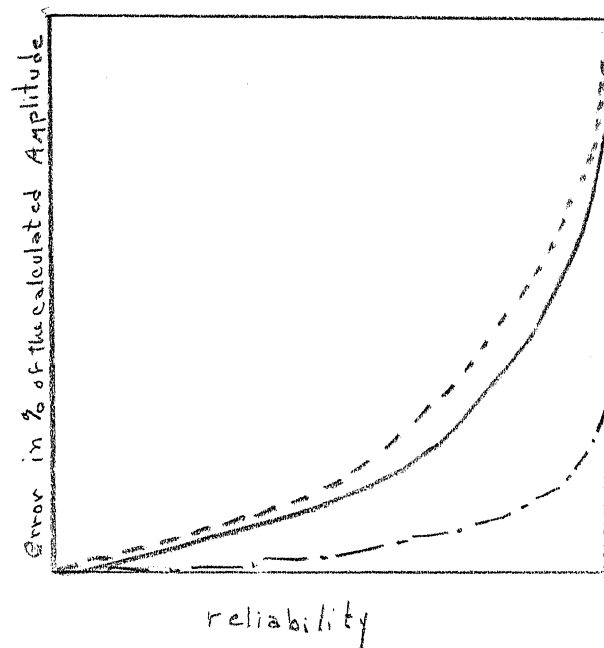


Figure 24. The reliability of the forecasting method for mean daily levels. 1-natural, 2-inertia, 3-of the method.

Experience in applying this equation in everyday forecasting has shown that it has a satisfactory justification and is actually very useful in forecasting. Similar relationships are formulated in the Marine Section of the Central Institute of Forecasting for many points in the Sea of Okhotsk and Far Eastern Waters.

In constructing relationships for areas where ebb-flow variations play an essential part the correct periodic changes in level precalculated by a harmonic analysis method are computed from the observations, and the difference found is developed in the usual manner used for seas without flood tides.

Predicted values of nonperiodic changes in level in seas with flood tides is algebraically added to the figures printed in the flood tide tables. Therefore, we must always bring the predicted level to zero on the chart before releasing a forecast.

The following should be especially noted. When the long-term variations in the level are great, as, for example, in the Caspian Sea, coefficients included in the equation used in forecasting must be recalculated when the mean annual level change is 30-50 centimeters, since the depth changes with a change in level. Theoretical studies, however, show that a change in the depth of a sea must be reflected in the value of sgonno-nagonnye variations.

The success achieved in constructing relationships for sgonno-nagonnye variations in level for several points in the Sea of Okhotsk gives us the opportunity to formulate several suppositions having an essential meaning in the study of this phenomenon in the future. It is well known that the depths in the Sea of Okhotsk are rather great,

as a result of which it would seem that the level variations there should be arrived at by means of the theory developed for a deep sea. But actually, the matter is different. If we examine the effective direction of the pressure gradient we will see that the running-up winds will be those which are directed onto the shore and not along the shore as we would expect for a deep sea. Evidently, the reason for these phenomena lies in the comparatively small level deviations from a position of equilibrium in the case of wind action over a deep sea. An examination of the physiogeographical conditions in which the above-named areas of the Sea of Okhotsk are found and for which equations were obtained that correlate the level deviation from a position of equilibrium with corresponding pressure gradients, shows that all areas have markedly shallow waters. One such is the Amur estuary, the Shantar Sea with the Bay of Ayan, and others. This indicates apparently a weak dependence of level changes on wind action over the deep areas of the sea; level variations are determined by wind action over the shallow sections of the sea. To study this problem we must observe currents and level variations simultaneously. Without studying current speeds and directions arising in the sea as a result of wind action some suppositions can be propounded or theories be made, but the solution to the problem can be only as a result of direct observations of currents. The observations of currents, however, in an open sea in general, and in the Sea of Okhotsk in particular, has many difficulties. The Sea of Okhotsk is characterized by highly developed ebb-flow variations in level and currents, and by drift currents difficult to study.

14. AN EXAMPLE OF THE COMPUTATIONS FOR WORKING OUT LEVEL FORECASTS

On a special blank we write the data of standard observations of air pressure at points located on a line coinciding with the effective direction of a pressure gradient. In the zero column we indicate the year, month and day. The pressure for each period corresponding to the dates is written in separate columns. If we bear in mind the four-period observations, then in column 5 we must write mean daily pressure values for each station. In this manner 10 columns will be filled with standard pressure observations and daily means for each station. In column 11 is written the difference in mean daily pressure always subtracting the pressure at the second station from that at the first. The pressure difference is always written with its sign. Column 12 is used to write in the sliding mean 10-day values of pressure differences. These differences, mean for a 10-day period, are obtained by a simple algebraic sum of the first 10 values written in column 11 and by a division of the obtained sum to one decimal place. The first figure appears in column 12 on the tenth line counting from the top down. To get the succeeding figures in column 12 it is enough to subtract algebraically from the figure on the tenth line of this column (having previously dropped the decimal point), the figure on the first line of column 11, and to add algebraically ~~the figure on the eleventh line of column 11~~ to the result, and again to divide the resultant figure to one decimal place. In computing the succeeding sliding mean 10 day differences we continue similarly.

The next five columns of the table are used for writing in

the standard observations of sea level and the mean daily level. In column 18 we have the sliding mean 10-day sea levels. They are computed exactly as the mean 10-day sliding pressure differences. Column 19 is filled in with differences between the last standard observation of a level found on line ten of column 16 and the sliding mean 10-day level found on the same line of column 18. Always we subtract algebraically the 10-day level from the standard level and the resultant is written with the appropriate sign. The difference $H_{\text{stand. obs.}} - \bar{H}$ is multiplied by k_1 and is written in column 20 on the same line.

In columns 21 and 22 we have the pressure forecast for 12 hours in advance. In column 21 we write the forecast for the first station, and in 22 that for the second station. Column 23 shows the difference: from the predicted pressure for the first station we subtract the pressure predicted for the second station. The subtraction is made algebraically, hence the difference is always written with its sign.

In column 24 we write the algebraic difference between the figures in column 23, $\Delta \bar{P}_{\text{forecast}}$, and the figures in column 12, $\bar{\Delta P}$. In addition to this the figures in column 12 are subtracted from the figures in column 23. The resultant is written with the appropriate sign.

In column 25 we write the figures of column 23 multiplied by the coefficient, in column 26 the algebraic sum of the figures on the same line of column 18, 20 and 25. The figure obtained in this manner will be the forecast of the sea level for the next 24 hours. In the computations it is necessary to watch in what units the pres-

sure and level must be taken.

In column 27 we write the actual level for the 24 hours for which the forecast was made, and in column 28 we indicate the difference between the actual and the computed levels.

15. FORECASTS OF CURRENTS

In working out a current forecast procedure we must determine, as in the case of forecasts of nonperiodic level variations, the relationships, not directly between the current and the wind, but between the current and atmospheric pressure. The relationship with the atmospheric pressure has the advantage, that the pressure is predicted, observed, and interpolated more simply than the wind. The forecast dependences which must be constructed in the immediate future, intended for current forecasts must contain, besides atmospheric pressure, the levels at respective points as well. In accordance with the above-mentioned conclusions obtained from theoretical studies, the pattern of currents is always closely related to the tide pattern. Inasmuch as it was possible to formulate a procedure for forecasting level variations, though not yet complete, we have every basis for believing the formulation of a forecast procedure for currents to be possible. Of course, at first only the main peculiarities in a development of currents dependent on a wind can be predicted. In forecasting currents the dependences must have a more complicated form than in forecasting tides. In forecasting sea level variations in many cases it was possible to disregard the transverse inequality of the wind and replace a detailed characterization of the wind pattern with rather rough mean values or even with pressure differences at two separate points. In forecasting currents

average characteristics will be insufficient, great detail will be necessary either of the wind field, or the air pressure field related thereto or related to these elements of sea level. The inclusion sea level in the corresponding dependences is advantageous also because it can be quite simply and accurately observed.

For the present, besides the above-mentioned attempt at establishing dependences between wind and current, we need mention only one effort which made it possible to predict nonperiodic currents in the Kerch Strait. This work was completed in the Marine Section of the Central Forecasting Institute by K. P. Vasil'ev.

The Kerch Strait was taken as the subject of research for several reasons: the existence of satisfactory products of observation, the comparative simplicity in the current pattern in the Strait, like that in a canal linking two seas, the need of this research in fulfilling the requirements of the national economy. To illustrate in what measure this problem is pressing, it is not uninteresting to enumerate certain facts. In spring time ice drifts from the Sea of Azov into the Black Sea are often observed. To forecast ice drifts, current forecasts are necessary. The movement of anchovies depends on the development of Black Sea or Azov currents through this Strait, and consequently, so does the success of the fishing seasons. Finally, the currents of any direction in the Strait determine the possibility of forming a stationary ice coat in the Strait, since even when a stationary ice coat is formed in the Strait it is sufficient for the Black Sea current to arise and soften this ice because of profuse melting top and bottom.

The difference in level between sections of the Black and

Azov Sea directly adjacent to the Kerch Strait is the direct reason for the changes and varied intensities of these currents. Short-term variations in sea level in these regions occur as a result of sgonno-nagonnye phenomena. It is essential in this case to combine the problem of forecasting currents in the Kerch Strait with the problem of forecasting level variations. Of course, it would be impractical to formulate dependences separately for forecasting level changes in sections of the Azov and Black Seas adjacent to the Kerch Strait, and from these predicted levels to determine the current slope and speeds. Such a course would create many errors. Therefore, Vasil'ev took the course of searching for the effective direction of the pressure gradient determining the current field directly in the Strait, and of establishing the dependence between the pressure difference taken between two definite points on a line coinciding with this gradient direction, and the projection of the current speeds on the axis of the strait. His research gave entirely satisfactory results, allowing the dependence he obtained to be used operationally in forecasting current directions with sufficient reliability.

This example shows the possibility of constructing similar dependences for forecasting in several other areas, for example in the Nevel'skiy Strait, the estuary of the Amur River, or La Peronse Strait etc.

Similarly dependences useful in actual forecasting can be obtained based on pressure or wind forecasts. For this purpose it is necessary to use first of all the observations of currents, without which at the present time the development of a forecast procedure for currents would be unthinkable.

CHAPTER V

FORECASTS OF SEA DISTURBANCES16. SOME FACTS ABOUT SEA WAVESPreliminary Comments

From the previous discussion it follows that air moving over the surface of a sea causes complicated systems of currents. Together with this, changes in the surface level of a sea occur which above were called sgonno-nagonnye variations. Together with the development of sgonno-nagonnye variations in level, free variations in sea level are developed with a period of about 12 and 24 hours. If the period of free variations is close to the period of the driving force, as, for example, in the above noted case of breeze winds, then these variations constantly supported by periodically shifting winds should be considered as being forced, capable of existing as long as the driving force acts. These variations will occur with the period of the driving force. For breeze winds the period of these variations is equal to 24 hours, for monsoon winds it is one year.

But the air motion over the water surface, aside from the currents and the variations in sea level over a long period connected with them, creates variations in the level surface of the sea of considerably smaller periods, measured in seconds. In this case we are speaking of wind disturbances in which the changes in the sea surface are quite noticeable. The agitated surface becomes uneven. The raised and lowered portions of the sea are clearly observed in com-

parison with the nonagitated surface of the sea. These irregularities on the sea surface can be observed only as a result of a displacement of water portions, and consequently, currents of a sort are also related to the disturbances, that is, the motion of water portions, as in the case of sgonno-nagonnye phenomena. But in sgonno-nagonnye phenomena currents, displacement of water portions and level changes are easily noted. In disturbances, the deviation of the level sea surface from a condition of equilibrium can be seen and fixed, but it is extremely difficult to follow the displacement of water portions. These displacements are sooner guessed at than directly observed. However, it has been established by theoretical and laboratory studies that during disturbances water portions complete fluctuating movements in a vertical plane, in most cases almost circular, whereupon the less the radius of the circumference along which the portions move the further from the surface is the moving portion located in depth.

Usually the following elements are differentiated in waves: crests -- the greatest deviations of the level surface of the sea from a condition of equilibrium upwards; troughs the greatest deviations of the level surface of the sea from a condition of equilibrium downwards; height of a wave -- the vertical distance between the ~~greatest deviations of the level surface from a condition of equilibrium~~, or the vertical distance between the trough and crest of a wave; wave length -- horizontal distance between two crests or between two troughs; period of a wave - the time in which two successive crests or two successive troughs pass through a stationary point; speed of wave -- the distance which a crest or a trough of a wave travels in a horizontal direction for a unit of time; steepness

of a wave -- expressed approximately by the relation of the wave height to its length.

The above enumerated wave elements are related to one another by strictly mathematical relationships. If we use the designations: h = wave height, a = wave amplitude, L = wave length, T = wave period, C = wave speed, $S = \frac{h}{L}$ = wave steepness, g = gravity acceleration, π = relation of the length of the circumference to its diameter; then we can write the following most simple relationships:

$$C = \frac{L}{T} = \sqrt{\frac{g}{2\pi} L} = \frac{g}{2\pi} T,$$

$$L = \frac{2\pi}{g} C^2 = \frac{g}{2\pi} T^2,$$

$$T = \sqrt{\frac{2\pi}{g} L} = \frac{2\pi}{g} C.$$

All these relationships are as true for one wave as for a series of completely similar waves -- in those cases when the depth of the sea is greater than half the wave length. However, if we glance at the sea surface at a time when disturbances caused by a wind are well developed, then immediately the question arises: what heights, lengths, periods, steepnesses and speeds. At sea during a disturbance developed by a wind, many waves exist at the same time. We can observe gigantic mountain-like waves, whose height is greater than 10 meters, and whose lengths measures tens of meters, simultaneously with small newly formed waves with a height measured in centimeters. That is why we must first decide which waves

should be studied and which should be carefully analyzed in order to predict these waves in the future.

As is well known, one of the methods of observing disturbances is that of stereophotography. This method permits determination of the condition of the agitated surface at some areas of a sea for a particular moment of time. Let us suppose that we have available several pictures of the sea surface made, for example, in the course of 5 minutes. After carefully examining these pictures we can form a table which will characterize the sea condition in this area during, say, a period of one hour. In the table we can include the wave heights and lengths determined on each picture.

Using the equation $T = \sqrt{\frac{2\pi}{g}} L$, we can complete this table with the computed values of wave periods, and then, with the aid of the equation $C = \frac{L}{T}$, the wave speeds. Having divided, in addition, the wave heights by their lengths in the equation $\xi = \frac{h}{L}$, the wave steepness can be easily found. In this manner the correct characteristics of an agitated sea surface can be derived. But when we limit ourselves to such a table it is very difficult to imagine the entire picture of the phenomenon which concerns us. To get more details on the sea condition it is necessary to subject this table to further development.

We can consider height the basic criterion characteristic of the wave. Accepting this condition we will break up the entire wave gamma enforced on the table into groups according to height: 0-0.5, 0.5-1.0, 1.0-1.5 meters etc. up to the maximum wave height observed and carried in the basic table. Next we must count the number of waves in each of these groups. After this we count the number of

waves in the following groups: 9-8.5, 9-8.0, 9-7.5, 9-7.0 meters etc. to 9-0 meters (here we assume that the highest wave was exactly 9 meters). Taking the number of waves in the group 9-0 as 100 percent, we compute the percentage of the general number of waves contained in the first, second, etc. groups.

After completing all these computations we construct a curve of wave height reliability. These curves usually resemble curves shown in Figure 25. In this figure taken from L. F. Titov's work we have several reliability curves corresponding to various degrees of disturbance given in balls.

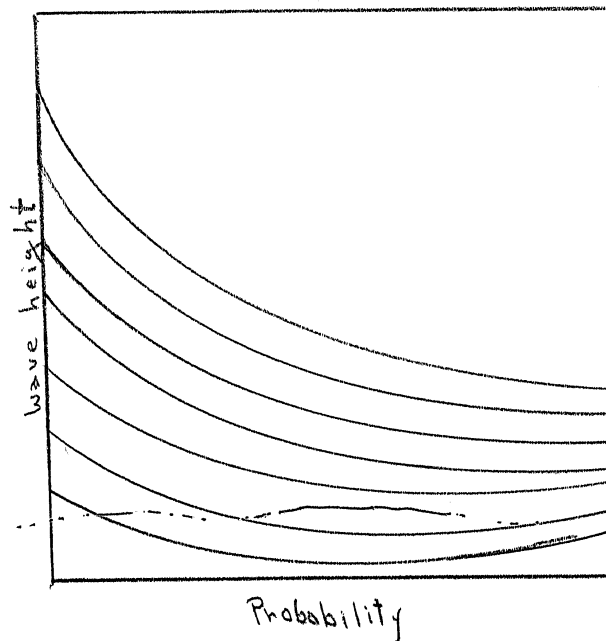


Figure 25. Probable wave heights for various degrees of disturbance.

The Roman numerals indicate the degree of disturbance in walls.

To get the reliability curves shown in Figure 25 the computations mentioned here were made. These reliability curves do not have very sharp variations because in constructing them a large number of observations were used. Inasmuch as the wave lengths are also values observed in stereophotographs, it is useful to construct a reliability curve for wave lengths in a similar way. The other wave elements can be computed by the formulas shown above.

Using such a reliability curve we are easily convinced that the large waves which usually catch the observer's eye compose a small percentage of the general number of waves. Nevertheless these same large waves have a significance in navigation, harbor construction and in shore erosion and so forth. It would be exceptionally important for research purposes to use just such reliability curves of waves in order to analyze the entire wave gamma observed at sea. But, unfortunately, a stereophotograph and its interpretations are too complicated and laborious to use this means of observation at ordinary hydrometeorological stations. Until the present time there are very few observations of wave elements by even the most usual instruments. The observer carefully attempting to complete his disturbance observations fixes his attention on the largest waves. When in the observations of wind disturbances the relative wave sizes are not especially limited, then this means that large waves are being watched. In the reliability curves constructed in the above indicated manner will be placed in the limits between 0-10 percent.

Otherwise there are swells. In this case the differentiation of waves is not as great and hence the necessity of discussing the highest or longest waves is lost. Waves and swells and especi-

ally "dead" swells greatly resemble each other.

In current instructions a swell is identified as a wave system remaining from a wind which died down or changed its direction up to the time of the observation, or as a wave system created by a wind, blown or blowing in an area removed from the area of observation.

In making observations doubt can frequently arise in evaluating swells, since it is uncertain how much the wind should die down or how much the direction of the disturbance should differ from the wind direction for the observer to call a disturbance a swell. It is not useful to recommend to the observer any rules by which he could automatically distinguish a swell from a wind disturbance. The user of the products of observations in working out forecasts must always be critical towards the reports received from stations and compare them with those obtained by his application of relations and with the reports of the general sea condition. He himself, for the purpose of the forecast, must decide on the correctness of data received from an observer. More precisely a swell should be considered as a subsiding disturbance. From this it does not follow at all, of course, that at the observation point there may not be increasing swells and that swells necessarily must subside from one period to another. On the contrary, quite frequently a swell may serve as a forerunner of a storm. This occurs when a swell formed out of a wind disturbance moves in the same direction as the barometric evolution that caused the storm wind. Wind waves, and especially a swell, frequently spread considerably faster than the barometric evolution.

Waves having a large period spread at a great speed

$$C = \frac{g}{2\pi} T$$

At the same time, waves possessing large periods have a large length. This circumstance has an very great significance for studying sea waves and for predicting them. To understand this we must remember and clearly note the process of the birth and death of waves.

17. THE DEVELOPMENT OF WAVES

Under wind action waves of small height, small length, and short periods, are first formed on a sea surface. Ripples, forming on a sea surface at the very beginning of a wind action, are nothing more than capillary waves. The basic forces responsible for the birth and development of capillary waves are uneven wind pressure, due to turbulence, and surface tension, acting at the border between the wind and air.

It is a distinctive trait of capillary waters that surface tension plays a significantly bigger part than gravity. These waves have a great steepness, up to 30 degrees, and their speed of spreading, in contrast to other wind waves is inversely proportional to the wave length. These waves are born before the others and are noted on a sea surface during the entire time that the wind action continues. ~~They interfere~~ and die down because of friction, but generally contribute to bringing ever more masses of water into motion. The wave lengths become greater, and the wave height also grows, but their steepness becomes considerably less.

As was mentioned above, on the sea surface after the passage of some time from the moment of wind development we can observe an

entire gamma of waves with exceedingly different periods, heights, lengths, and steepnesses. All these waves mix in with one another, that is, fold in together so that as a result the wave build-up becomes increasingly larger. The larger the wave period and length, the more alive seems the wave, and the greater the speed with which the wave spreads. That is why only waves having great lengths and periods get through from storm areas to regions of relatively little wind as swells. Swells are always of greater length and longer period than the waves from which they are formed.

Inasmuch as waves and swells have different periods, the longest waves come first from a storm area point located sometimes thousands of kilometers from the storm area. True, these waves often have such small amplitudes and, in the main appear so slanting, that it is difficult to catch them with the naked eye. However, it is extremely important to follow the appearance of such swells, as they may shortly -- in one, two, or even three days -- give warning of an approaching storm or the main "front" of the swells. This condition has particular importance when forecasting storms, wind disturbances and swells. To trace these waves special equipment placed at great depths (75-100 meters has been brought into use recently. Because this apparatus is installed at great depths it does not record waves of short periods at all, which to a considerable degree simplifies the analysis of the recordings of this equipment. The special arrangement which the apparatus provides also excludes recording variations of too large a period such as ebb tides and sgonno-nagonnye phenomena. Nonetheless, the analysis of the recordings of this equipment is so complicated that another special apparatus is required -- an analyzer, which permits us to record ac-

curately the time of approach of a forerunner of swells and its period. An average period of a swell in an ocean is 10 seconds. With this period the speed of travel for the swell would be about 44 nautical miles in one day, and the period of a forerunner of swells was recorded at 30 seconds. The group speed corresponding to this period is about 1000-1100 nautical miles per day. A detailed explanation of the term "group speed" will be given below. If an uninterrupted record is made of the period of a forerunner of swells and a steady decrease in the period is noted, then a qualitative conclusion can be made that the storm disturbance area or the source where the waves of the forerunner of swells are born are approaching the observation point. The farther the waves move from the point of their birth the smaller their amplitude and the larger the wave length, and, consequently, the larger the period. This remarkable peculiarity of the spreading of swells permits us to construct two formulas with whose help we can compute the distance which the swells travelled from their place of birth to the observation point as well as the time taken for this movement. These formulas have the following form:

$$S = \frac{gT^2}{4\pi\left(-\frac{\Delta T}{\Delta t}\right)}, \quad t = \frac{T}{\left(-\frac{\Delta T}{\Delta t}\right)}.$$

Here S = path travelled by the wave from the source to the point of observation; t = time during which the wave travelled from the source of the wave's birth to the point of observation; T = period characteristic of a group of forerunners of waves; ΔT = difference between the periods fixed by two successive observations; Δt = time difference between two successive observations of periods; g = gravity

acceleration: π - ratio of the length of circumference to its diameter, equal to 3.14.

In order to use effectively the forerunner of swells it is necessary to make the observations and analyze them systematically, and compare the calculations from the above mentioned formulas with the synoptic weather maps.

A concomitant analysis of synoptic charts with data from wave stations can be extremely effective in predicting not only a wind disturbance but also swells.

18. THE TRANSFER OF WAVE ENERGY

To forecast disturbances and swells successfully it is necessary first of all, to have a picture of the transfer of energy for the medium in which they are spreading. It is easiest to determine this with the aid of an analysis of the advance of a single wave or a series of entirely similar waves into a quiet calm water surface. If we throw a stone into the center of a round body of water and watch the waves spread, then it is easy to note that the waves are very clear in the center, not far from the center the water surface hardly deviates from a position of equilibrium, while at the edge of the basin this disturbance may remain entirely unnoticeable.

The full wave energy always consists of two equal parts -- kinetic energy determined by the motion of masses, and potential energy determined by that action which is completed in a rise of water masses over the undisturbed level.

Kinetic energy can be expressed by the relationship:

$$W_k = \frac{1}{16} g \rho h^2 = \frac{1}{4} g \rho a^2$$

Potential energy is expressed in the exact same way

$$W_p = \frac{1}{16} g \rho h^2 = \frac{1}{4} g \rho a^2$$

and the full wave energy in relation to the above mentioned equality of kinetic and potential energies can be written thus:

$$W = \frac{1}{8} g \rho h^2 = \frac{1}{2} g \rho a^2$$

In these formulas g = gravity acceleration; ρ = water density; a = wave amplitude; h = wave height.

Hence wave energy is expressed for each of the disturbed surfaces.

Let us suppose that we have a long narrow channel on the end of which a float is placed. With the help of a special rig we can make this float complete correct variations and strike the water surface in the channel at even intervals of time. As a result of the harmonic movements of the float waves will run along the channel.

Studying the potential and kinetic energy of each wave just formed by the blows of the float we must note that the greatest values of potential energy will be attained at those points of a wave where the largest deviations of the water surface from a position of equilibrium will be observed; the lowest values of potential energy or, to be more exact, a zero value will be at the point where the deviation of the water surface from a position of equilibrium is equal to zero, or where the wave surface coincides with

the undisturbed surface. The change in potential energy along the wave profile is shown in Figure 26. As we see from this figure the potential energy is the periodic function changing in time and expanse together with the wave form.

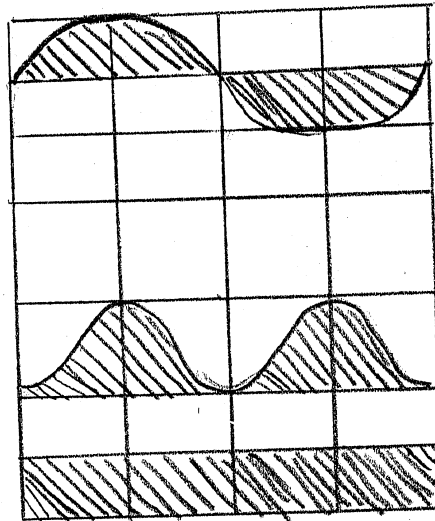


Figure 26. Wave energy.

Kinetic energy, related to the movement of water masses, is equally distributed along the entire wave and is independent of the deformation of a sea surface (Figure 26).

How then, in that case, does the energy spread when a wave moves to an area where there is no disturbance. Apparently, the static condition of water masses is responsible for the undisturbed sea surface. If, now, the first wave agitated by our float moves one wave length in the direction of the undisturbed water surface then this is unavoidably connected to a movement of water masses with definite speeds. An energy proportional to this speed can be obtained only on account of the potential energy moving together with

the wave form. As the full energy of any wave always consists of equal values of kinetic and potential energy, half the potential wave energy, having moved ahead into a region of an undisturbed wave, must be expended on bringing stationary water masses into motion. But the expenditure of potential energy related to the wave height will decrease the wave height. Having travelled a distance of one wave length in the direction of the undisturbed surface, the wave which we are studying also expends its potential energy on moving undisturbed water masses and, consequently, will lose its height again. This will continue until the wave height moving continuously forward appears unnoticeable to the observer.

The second wave agitated by the float moves forward behind the first and therefore will encounter on its way the water masses already in a moving condition, since they were set in motion by the first wave. Therefore, the second wave loses less potential energy in converting it to kinetic, and consequently the decrease in height of the second wave will be less than was the case in the first wave. The same occurs with the third, fourth, etc. waves. In this manner the second wave will be noticeable at a greater distance from the float and the third at still greater distance, etc. Let us suppose that the float stops operating. Then, it would seem, we could expect the group of waves agitated by the float to end with the last wave caused by the float and to move ahead in the direction of the undisturbed water surface, and after the wave caused by the last blow an undisturbed and flat water surface will be restored again. But actually this is not so. The energy of the last blow of the float, just as for all the previous ones, was divided into two equal parts; one half was expended in changing the appearance of the water

surface, and the other half was expended in moving water masses. Consequently, in the case of the last blow half of the energy creating the form, that is, the potential energy, was transferred forward as in the case of all the previous waves, but the kinetic energy was unexpended. Therefore, the unexpended kinetic energy related to the last blow of the float will change to potential and behind the wave caused by the last blow another wave will appear.

In this manner an entire series of waves will be formed with each blow of the float, the heights of which gradually decrease with the distance from the float in the direction of the undisturbed water surface. These series are superimposed one on the other. As a result of the combined series of waves a complicated picture is obtained of the dispersion of wave heights and accordingly of the energy of wave motion. Let us watch now how the wave height and its energy W will change.

The first blow of the float causes a wave with energy $\frac{1}{2}W$. Up to the time of the second blow the first wave will travel one wave length and in doing so will lose (leave) $\frac{1}{4}W$ at the place where the second wave will form, that is, at the float. The second blow of the float will again transfer to the water surface an energy equal to $\frac{1}{2}W$. This energy, the energy of the second blow of the float, will combine with the remaining energy from the first wave, that is, $\frac{1}{4}W + \frac{1}{2}W = \frac{3}{4}W$. Half the energy formed as a result of the combined waves, that is, $\frac{3}{8}W$, again will travel from the float, then $\frac{3}{8}W$ will also remain at the float. The third blow will again transfer to the water surface $\frac{1}{2}W = \frac{4}{8}W$. This energy of the third blow will again combine with the remaining energy at the float $\frac{3}{8}W + \frac{4}{8}W = \frac{7}{8}W$. This will continue after each blow. The energy

transferred to the water surface by the float will be combined with the energy remaining after the previous blow. The wave nearest the float after the n -th blow will have the energy $\frac{2^n - 1}{2^n} W$. The wave that succeeds in moving the greatest distance from the float will contain the least amount of energy as compared with all the other waves, to wit: $\frac{1}{2}W$. The wave located half way between the wave nearest the float and the wave farthest away will have an energy equal to $\frac{1}{2}W$, that is, just the amount of energy which is transferred by the float to the water surface. In Table 5 the disposition of wave energy in the channel after 10 blows of the float against the water surface is shown as an illustration [See Table 5 on following pages].

From examining Table 5 it follows that a wave having travelled 10 lengths has an infinitely small energy -- totaling only $1/1024 W$. The series of figures located diagonally represents the energy loss of one wave travelling along an undisturbed water surface. The figures in column 1 represent the energy of the wave formed directly by the float. Columns 2, 3, etc. pertain to the second, third waves counting from the float, etc. An outstanding fact is that on the lines representing the odd series of waves or blows of the float the middle figure on the line is always $\frac{1}{2}W$. This confirms the transfer of energy of groups of waves or, as we say, the group wave, with a speed equal to half the speed with which each wave of the group travels.

On the lines having even numbers this cannot be noted in the table from the figures there, since $\frac{1}{2}W$ is exactly at half the distance travelled by the first wave moving along an undisturbed water

TABLE 5

No of flows of the float

resulting in a series of waves Value of wave lengths, occurring in the channel

	1	2	3	4	5	6	7	8	9	10
[1]	[2]	[3]	[4]	[5]	[6]	[7]	[8]	[9]	[10]	[11]
	$\frac{1}{2}$									
1	$\frac{3}{4}$	$\frac{1}{4}$								
2										
3	$\frac{7}{8}$	$\frac{4}{8}$	$\frac{1}{8}$							
4	$\frac{15}{16}$	$\frac{11}{16}$	$\frac{5}{16}$	$\frac{1}{16}$						
5	$\frac{31}{32}$	$\frac{26}{32}$	$\frac{16}{32}$	$\frac{6}{32}$	$\frac{1}{32}$					
6	$\frac{63}{64}$	$\frac{57}{64}$	$\frac{42}{64}$	$\frac{22}{64}$	$\frac{7}{64}$	$\frac{1}{64}$				
7	$\frac{127}{128}$	$\frac{120}{128}$	$\frac{99}{128}$	$\frac{64}{128}$	$\frac{29}{128}$	$\frac{8}{128}$	$\frac{1}{128}$			
8	$\frac{255}{256}$	$\frac{247}{256}$	$\frac{219}{256}$	$\frac{163}{256}$	$\frac{93}{256}$	$\frac{37}{256}$	$\frac{9}{256}$	$\frac{1}{256}$		
9	$\frac{511}{512}$	$\frac{502}{512}$	$\frac{466}{512}$	$\frac{382}{512}$	$\frac{256}{512}$	$\frac{130}{512}$	$\frac{46}{512}$	$\frac{10}{512}$	$\frac{1}{512}$	
10	$\frac{1023}{1024}$	$\frac{1013}{1024}$	$\frac{968}{1024}$	$\frac{848}{1024}$	$\frac{638}{1024}$	$\frac{386}{1024}$	$\frac{176}{1024}$	$\frac{56}{1024}$	$\frac{11}{1024}$	$\frac{1}{1024}$

surface. In other words, we can write it as $m = \frac{n-1}{2}$,

where m is the distance, expressed in wave lengths, and n is the number of time the float struck or the series of waves created by a blow. Consequently, if n is even, then m is fractional; but in the table the figures used are only for full waves.

We must note that the energy of a group of waves having travelled more than half the distance between the first wave and the float is infinitesimal. The energy of waves at the half way mark grows sharply and quickly approaches the maximum after the middle wave. On Figure 27 a case is presented of the change in wave energy by measuring its pressures from the point of origin. For the case indicated on the figure, n is taken as exactly 899, and consequently, m is equal to 450 wave lengths. The vertical scale on the left of figure 27 indicates the relationship of the wave height to the greatest height, expressed in percent, and the right scale indicates the percent of the maximum energy. On Figure 27 the middle wave is exactly at the half way mark of the path of the first wave. The energy of the middle wave is equal to half the maximum energy and its height is close to 70 percent of the maximum height. A sharp increase in wave height takes place near the middle wave. The wave height increases from 10 to 90 percent at a distance of 50 wave lengths. [See Figure 27 on following page]

A simplified example was studied above in which waves arising from a harmonic variation of the float flowed out in a straight line channel. These waves had the same period, length and speed. In nature, as was already mentioned above, waves arising on a sea surface are developed by wind action and represent

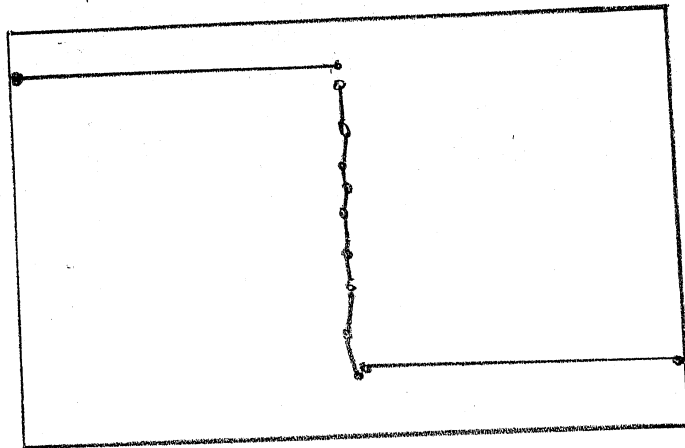


Figure 27. The change in wave height relative to the distance from the point of origin when a wave spreads along a calm (not agitated) sea surface.

On the horizontal scale the distance expressed in wave lengths is given. On the left vertical scale the percent of the maximum wave energy is given -- W/W_1 100, and on the right the percent of the maximum wave height. The big circles indicate the waves; at the point of origin, the middle and very first wave leaving the point of origin.

an entire gamma of periods, lengths, and speeds. Nonetheless, the example studied above allowed a very important relationship between the speed of a single wave and the group speed to be established. It also shows that the height of the group wave spreading in the direction of an undisturbed sea surface is infinitely small for the leading waves of the group and only at the half way mark of the

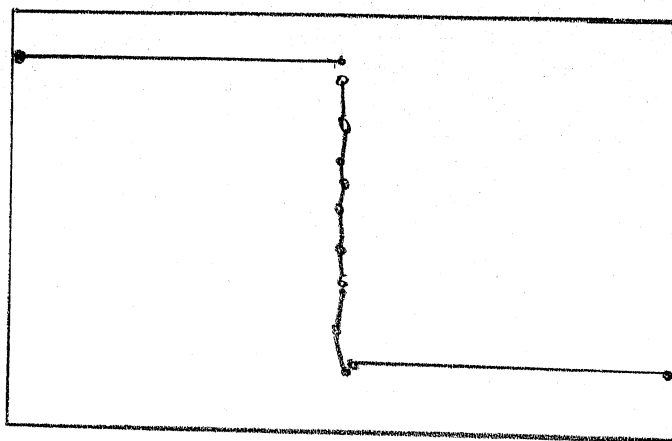


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an entire gamma of periods, lengths, and speeds. Nonetheless, the example studied above allowed a very important relationship between the speed of a single wave and the group speed to be established. It also shows that the height of the group wave spreading in the direction of an undisturbed sea surface is infinitely small for the leading waves of the group and only at the half way mark of the

path travelled by the leading waves does the height of the group wave suddenly increase and quickly reach maximum values. This phenomenon has an extremely important practical significance in predicting disturbances or swells, inasmuch as at sea we must contend with entire gammas of waves.

In computing elements of group waves it is best to take the wave speed, period, height and length, averaged for large waves, limited on the reliability curve to 0 and 10 percents. The group wave speed should be studied as half the mean speed of large waves. The leading waves, however, may often serve as wave-forerunners, consequently, the discovery of waves of small amplitude but large period can frequently assist in predicting either the approach of heavy swells, or the approach of a storm.

19. WIND ENERGY AND WAVE ENERGY

Predicting disturbances or swells is greatly complicated by the fact that not one of the elements remain constant, unchanged neither in time nor in space.

A wind disturbance is created by wind action. The influence of the wind on a water surface can be represented by the action of two component wind forces -- the normal wind pressure on a water surface and the tangential pressure. The normal wind pressure on a water surface transfers the wind energy to the sea surface. The mean wind energy value transferred to the sea surface by the normal pressure R_N , can be computed by the relationship:

$$R_N = \pm \frac{1}{2} \sigma \rho' (V-C)^2 A^2 a^2 C,$$

here σ_{λ} is some coefficient of proportionality; ρ' is the air density; V is the wind speed at the height of observation over the sea surface; C is the wave speed; $k = \frac{2\pi}{L}$, where L is the wave length; a is the amplitude (half the height) of the wave.

A plus sign (+) refers to a case when the wave speed is less than the wind speed, and a minus sign (-) is used when the wave speed exceeds the wind speed. Consequently, a wave receives energy from a normal wind pressure only when the wind speed exceeds the wave speed. If, however, the wind speed is less than the wave speed, the wave then loses energy and subsides. To understand the reason for this it is enough to recall the appearance of an agitated sea surface on which the wind acts.

The windward side of wave crests is more slanting than the side in whose direction the wave is moving. One look at a sea surface covered with such waves shows that the wind presses on the rear portions of the crests and appears as though it changes the waves. If irregularities appear on a sea surface the air cannot travel over it in a laminar flow -- a whirlwind movement of air masses begins. As a result of this an uneven air pressure develops on the windward and lee sides of the crest.

If the wave speed is greater than the wind speed, then the pressure will be greater not on the rear side of the crest but on the forward side. As a result the wave motion will be slowed down.

Observations have shown that the minimum wind speed for which a wave will develop must be about 1.10 centimeters/second. At this

wind speed waves will be developed which will travel at a speed of 35 meters per second, having a length of 8 centimeters and a period of 0.22 seconds.

At wind speeds greater than 5 meters per second a tangential (adjacent) pressure noticeably develops on the sea surface, which we can express thus:

$$P = \gamma^2 \rho' V^2$$

where ρ' is the air density; V is the wind speed at the height of the observer; γ is the coefficient of resistance $\gamma^2 = 2.6 \times 10^{-3}$.

The energy transferred to the sea surface by the tangential pressure can be computed by the relationship:

$$R_T = \gamma^2 \pi^2 \rho' \delta^2 c v^2$$

here δ is the wave steepness.

Since the waves receive energy as a result of the normal and tangential wind pressures, it appears that they can develop even when the wave speed exceeds the wind speed. This occurs because the wind speed is practically always greater than the speed of the horizontal motion of water masses, and only in the opposite case would the energy received by a wave from a tangential wind pressure be negative.

In the final analysis the development of waves will be determined by the amount of energy received from the normal as well as the tangential pressure, but inasmuch as the normal pressure depends on the wind and wave speeds it is well to introduce a new concept of wave development, which regards wave development as a relationship of the wave speeds to the wind speed.

The designation is introduced because in the beginning of wave

developments they have a speed considerably less than the wind speed. Developing further, the wind and wave speeds become more equal, and, finally, it can be shown that the speed of waves will exceed that of wind.

When wind travels over a sea surface it continuously uses up or transplants energy. Energy is used up on turbulent motions, on forming waves on the sea surface and on the development of drifts.

All the energy used up by the wind in the lower 10-meter layer can be computed with the aid of the relationship:

$$R_v = \tau v = \gamma^2 \rho' v^3$$

The energy expended in creating waves by normal pressure R_N and by tangential pressure R_T is used up in changing the height of waves R_h and in increasing their speed R_c . On Figure 28 a general quantity of energy is shown as a percentage of the entire energy R_v , studied in the lower layers set in motion by the air, which goes into forming large waves:

$$\left(\frac{R_N}{R_v} + \frac{R_T}{R_v} \right) 100$$

and separately the amount of energy transmitted by the normal $\frac{R_N}{R_v} 100$ pressure (solid lines). On Figure 28 the dotted lines indicate the amount of energy expended in changing the height of waves $\frac{R_h}{R_v} 100$, and in changing their speed $\frac{R_c}{R_v} 100$.

On the horizontal axis the relation of the speed of waves to wind speeds is plotted. The values β " representing the greatest steepness of waves are plotted on Figure 28 in special markings, $\beta = \frac{c}{v} = 1$, that is, when the wind speed is equal to the wave speed.

and, finally, β_m is plotted when the waves attain maximum values.

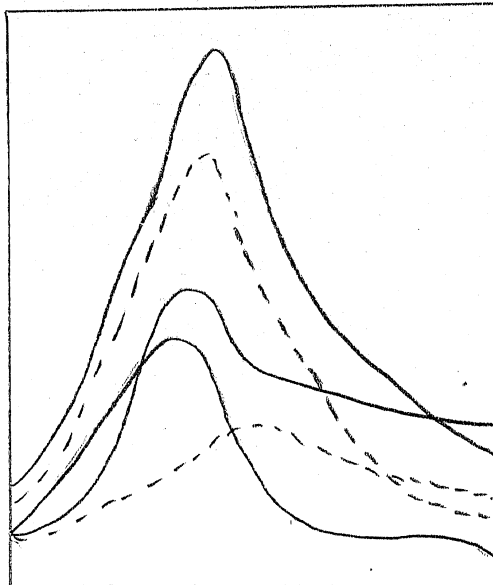


Figure 28. The gain and loss of energy of developing waves depending upon their age.

I -- the general amount of energy expended in forming large waves, expressed in percent of the entire energy transferred to the lower layer by a moving wind $\frac{R_N}{R_V} + \frac{R_T}{R_V} 100$; II -- the amount of energy transferred to the waves only by the tangential pressure

$\frac{R_T}{R_V} 100$; III -- the amount of energy transferred to the waves only by the normal pressure $\frac{R_N}{R_V} 100$; IV -- the amount of energy expended only in changing the height of waves $\frac{R_h}{R_V} 100$; V -- the amount of energy expended in changing the speed of waves $\frac{R_c}{R_V} 100$.

From Figure 28 it follows that the greatest comparative amount of energy is received by the waves at maximum steepness, whereupon

this energy acts mainly at the expense of the normal wind pressure. Furthermore, with an increase in wave speed relative to the wind speed a sharp decline takes place (relative to R_v) in the energy acting at the expense of the normal pressure, and, finally, when the wave speed reaches values equal to wind speed the energy received by the waves at the expense of the normal wind pressure decreases to zero, and for a further increase in the wave speed relative to the wind speed it becomes negative. Conversely, energy from the tangential wind pressure is everywhere positive and less than the energy received from the normal wind pressure only until the wave steepness reaches a maximum.

It is necessary to bear in mind that we are speaking in this case only of large waves predominating on a sea surface. As has been repeatedly noted, in the development of waves small sized waves are always observed together with large waves. These waves constantly receive energy from the wind in larger amounts than large waves do, because they possess a big steepness, and as a result of interference and other processes they transfer their energy to the big waves and subside themselves when the wind subsides.

The energy received by a wave from the wind is lost not only in changing the wave height, but also in changing other elements, partly in changing the speed of waves. On Figure 29 the amount of energy expended in changing the height and speed of waves is shown. As is apparent from Figure 29 a considerable amount of energy is expended in changing the amplitude of waves when they reach the maximum steepness (compared with Figure 30). After this the amount of energy transferred to R_v , expended in increasing the wave ampli-

tude, decreases together with an increase in the wave's age. When the wave's age reaches 0.8 the amount of energy expended in increasing the amplitude becomes equal to the amount of energy expended in increasing the wave speed. When a wave grows from 0.8 to 1.0, that is, when the wave speed becomes close to the wind speed and finally equals it, the amount of energy expended in increasing the wave amplitude and in increasing its speed is almost the same. A further increase in wave age causes greater and greater amounts of energy to be expended on its speed in comparison with the amount of energy expended in increasing the amplitude. It is necessary to understand that the results of the observations shown in Figure 28 refer only to large waves.

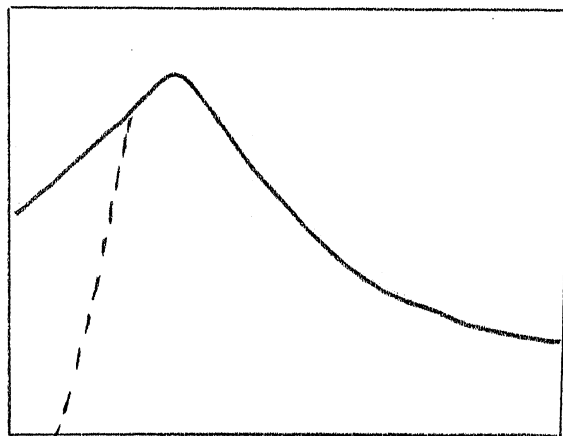


Figure 29. Relationship of a wave steepness $\frac{h}{100}$ to a wave

speed $\frac{C}{V}$.

[Figure 30 shown on following page]

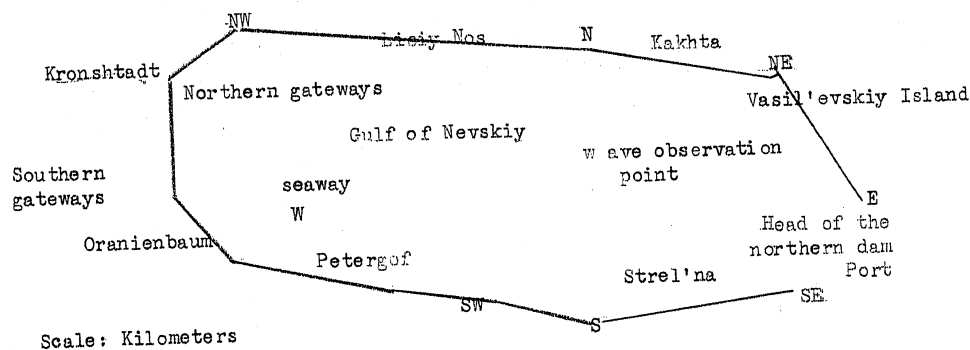


Figure 30. A rose of the wind dispersal in Gulf of Nevskiy.

20. THE RELATIONSHIP OF WAVE STEEPNESS TO AGE

Above, the value β " was mentioned as representing the maximum wave steepness. It is necessary to bear in mind that the steepness of waves is characterized by the relationship of a wave height to its length. It appears that the steepness of waves is closely related to another parameter, the wave's age. On Figure 29 a part-theoretical, part-empirical curve is shown comparing these two parameters. The points shown in this figure represent the observations made not only in the most varied regions of the ocean and sea but also in ponds. Consequently, the curve shown in Figure 29 comparing the wave steepness with its age can be considered sufficiently

universal.

Most of the observations show that the steepness of waves -- the ratio $\frac{h}{L}$ -- does not exceed 0.1 or 10 percent; theoretical studies, however, indicate the possibility of waves having this steepness. While the theoretical studies are not being supported by more dependable observations we must assume the maximum possible steepness as $1/10$. Thus, the maximum steepness observed for a wave's age is equal to 0.4. As we would expect from examining this curve at the beginning of a wind action the wave steepness increases as long as the wave's age $\frac{C}{V}$ does not reach a value of 0.4. Further wind action at first brings about a rather rapid decrease in the steepness and thereafter a slower one. Apparently, the length as well as the period and height of a wave never remain constant; they change together with the change in the wave's age.

21. THE DURATION OF WIND ACTION AND WIND RUN

Now, when the problem of the transfer of wind energy to waves and of the expenditure of energy show increasing the height of waves and their speed is clarified, it is easy to see that the longer the wind acts the more energy will be transferred to the waves. Besides, it is also apparant, that the larger the area over which the wind acts the more energy will be transferred to the waves. Besides, it is also apparent, that the larger the area over which the wind acts the more wind energy is transferred to the waves. Assuming that the wind has the same speed and direction over a large area, it is best to study not the area over which the wind acts, but the extent to which the wind direction coincides with the direction of waves. Of

course, the greater the straight line distance over which the wind acts over a disturbed sea surface the more energy can be transferred to the waves, and, consequently, other elements as well as the height of waves cannot be studied independently from this distance which is called a wind run.

Hence, we must add two more parameters to the parameters already studied on which the big waves depend: The continuance of wind action, taken for the time in which the wind acts with constant force and direction, and the wind run, taken for the straight line distance over which the wind acts with a steady force and direction on the waves. To make a clear presentation of the influence of these alternates on the development of waves it is sufficient to present the development of waves with the wind acting on a water surface blowing seaward and, on the other hand shoreward. Undoubtedly, in the former case the disturbance at the shore is considerably weaker than at sea. In the latter case, that is, when the wind action is shoreward, the disturbance is approximately the same in the open sea as in the offshore area (in both cases, of course, this is for an offshore area in which the disturbance is not distorted by shallow waters).

It is entirely apparent also, that a wind arising suddenly cannot create heavy disturbances. Some time is required for the wave elements to react to the wind force. It must be understood that limits exist for a wind run as well as a continuance of wind action beyond which ~~an increase in the run or the continuance of wind action~~ cannot be reflected in the wave development.

Summing up the above-mentioned development of waves at sea, it is important to note that in the long run the development of waves

is determined by the amount of energy transferred to a water surface by air currents moving over a sea.

The transfer of energy to a water surface depends on the following conditions: (1) on wind speed, (2) on the continuance of wind action, (3) on the length of the course over which the wind acts in a fixed direction on a sea surface, or on the wind run.

At some given wind speed the wave height grows very rapidly, at first, as do the speed of waves and their length. L. F. Titov's observations prove that when $C = 0.8$ the growth of wave height is almost stopped on account of the normal wind pressure. Increase in the speed of wave spread takes place mainly because of the tangential pressure. From the moment the wind speed becomes less than the speed of wave spread resulting from a normal wind pressure, the energy of waves decreases but due to the tangential wind pressure it continues to grow. In this connection we can presume without much error that the total energy of waves expended in increasing wave height neither decreases nor increases with the wave length and period. The wave length, speed and period continues to increase even when the wind stops entirely or when the waves exceed the limits of wind action. In the latter case the increase in speed occurs because of the decrease in wave height.

If the duration of wind action be designated by t_v ; the time required for the wave height to reach a maximum height corresponding to a given wind speed by t_h ; the time required for the wave speed to reach the wind speed by t_c , then the following conclusions may be made: when $t_v > t_c$ the continuance of wind action does not limit the wave development. When $t_v < t_c$ the waves of a given wind speed will not reach the maximum possible values of speed, length and per-

iod; when $t_v < t_h$ the waves will not reach even the maximum height.

If we similarly designate D_v as the length of a wind run; ^(D_c) as the length of the wind run when the wave speed can reach a given wind speed; D_h as the length of the wind run when the wave height for a given wind speed can reach the greatest value, then when $D_v > D_c$ the length of the run does not limit the development of waves for a given wind speed; when $D_v < D_c$ the disturbance for a given wind speed will not reach the development of speed, length and period corresponding to this wind speed; when $D_v < D_h$ the waves will not even reach the maximum height possible for the given wind speed.

L. F. Titov, assuming that the limit of height is reached by a wave when $C = 0.7$ or 0.8 , and using the empirical formulas constructed by him from a great many observations, worked out a very interesting and in practice very important table (Table 6). [See Table 6 on following pages].

With the help of Table 6 extremely interesting conclusions can be made. This table is applicable only for large waves. Consequently, large waves having reached a maximum height for a given wind speed, moving ahead and remaining under wind action of the same force, do not change their attitudes further on at the time when smaller waves continue to grow until they reach the maximum values possible for the given wind. As a result an increasing number of waves will prove to be of the same height. This condition allows us to examine the values shown in the table D_h and t_h as representing the beginning of the incorrect three-dimensional wind disturbance changing to the correct wind disturbance. In an entirely analogous manner we can examine the values D_c and t_c as the beginning

TABLE 6

balls	Wind	D_h		t_h	h	L_h	$\frac{h}{L_h}$	D_c	
	meters/seconds	kilometers	nautical miles	hours	meters	meters		kilometers	nautical miles
[1]	[2]	[3]	[4]	[5]	[6]	[7]	[8]	[9]	[10]
IV	6	57	31	4.6	1.1	13	$\frac{1}{12}$	211	113
V	9	134	72	7.2	2.2	30	$\frac{1}{14}$	364	197
VI	11	204	110	8.9	3.0	45	$\frac{1}{15}$	487	263
VII	14	333	182	11.7	4.6	72	$\frac{1}{15}$	698	377
VIII	17	509	275	14.5	6.4	108	$\frac{1}{17}$	947	511
IX	20	715	386	17.3	8.4	149	$\frac{1}{18}$	1230	664
X	23	969	523	20.2	10.7	197	$\frac{1}{19}$	1561	843
XI	27	1344	725	24.1	14.0	272	$\frac{1}{19}$	2039	1102
XII	30	1676	905	27.0	16.8	336	$\frac{1}{20}$	2448	1322

TABLE 6 (cont'd, right hand part

[11]	[12]	[13]
12.5	22	$\frac{1}{20}$
15.2	50	$\frac{1}{23}$
16.9	75	$\frac{1}{25}$
19.7	123	$\frac{1}{27}$
22.5	181	$\frac{1}{28}$
25.3	250	$\frac{1}{30}$
28.2	330	$\frac{1}{31}$
32.1	456	$\frac{1}{32}$
35.0	564	$\frac{1}{34}$

of the wind disturbance changing to a swell.

Actually, it is necessary to distinguish between the wind run and the wave run. If, above, the wind run was defined as the distance over which a wind acting over a water surface does not change its direction more than 35 degrees, then a wave run should be defined as the distance over which a disturbance spreads, also not changing its direction more than 35 degrees. Waves may spread outside the limits of wind action, therefore, the wave run D_B should be composed of a wind run D_C corresponding to a full development of all wave elements for a given wind force, and of the path over which a disturbance spreads, as a swell D_3 , that is;

$$D_B = D_C + D_3.$$

Whence it follows that after a wind wave reaches full development for distance D_C and time t_C , in the further progress of its development and in preserving a wind of previous force and direction, new peculiarities will be noted characteristic of swells, regardless of whether the wind is acting over a sea surface or has stopped. For these peculiarities we must first of all refer to the increase in the period length and wave speed.

22. COMPUTING WAVE ELEMENTS

Bearing in mind the above exposition and the mentioned empirical formulas, L. F. Titov constructed tables with the aid of which we can compute all the disturbance elements of an open sea for a given wind field. Since these tables have a wide scope only a small part of them is used here as an example (Table 7). In the full table,

however, we have altitudes, lengths, wave periods, wave growth continuance for wind speeds from 4 to 30 meters per second for various wind runs. On the underlined line of each table we have the wave dimensions which they attain for a given wave spread, that is, the underlined line is a repetition of the values in Table 6 for a given wind speed.

The wave heights shown in this table refer only to a large waves having a reliability or a frequency of about 5 percent. In this manner about 5 waves out of 100 may become higher than those determined by the use of this table. To change to waves having 1 percent frequency, that is, to almost the very highest, we must multiply the wave height and all the other elements taken from the table by 1.2. [See Table 7 on following pages]

When using these tables it is necessary to bear the following in mind:

- (1) the wind speed over the entire wind run is assumed to be constant,
- (2) wind direction does not change over 35 degrees,
- (3) the sea was calm over the entire wind run until the beginning of its action -- there was no disturbance,
- (4) the sea depth for the entire run is not less than half the wave length.

The data entered in the table are the speed, the direction of wind action and wind run. In the case when duration of wind action is large and the run is limited, the data corresponding to a given run is taken from the table. If, however, the duration of wind action is not large (and less than the value shown in the table

TABLE 7

Kilometers	Wind run		Wave Elements		Duration of wave
	Nautical	Height, Length, period,			growth, hours
	Miles	meters meters seconds			
[1]	[2]	[3]	[4]	[5]	[6]
Wind speed 20 meters/second					
20	11	1.2	15	3.1	1.5
40	22	1.8	23	3.8	2.4
60	32	2.2	30	4.4	3.1
80	43	2.6	35	4.8	3.9
100	54	2.8	42	5.2	4.5
120	65	3.2	47	5.5	5.1
140	76	3.4	52	5.8	5.7
160	86	3.7	57	6.1	6.2
180	97	3.9	61	6.3	6.8
200	108	4.2	65	6.5	7.3
220	120	4.4	70	6.7	7.8
240	130	4.6	73	6.9	8.2
260	140	4.8	77	7.1	8.7
280	151	5.0	81	7.2	9.1
300	162	5.2	85	7.4	9.5
400	216	6.0	102	8.1	11.7
500	270	6.8	118	8.7	13.6
600	324	7.6	132	9.2	15.4
700	378	8.2	146	9.7	17.1

TABLE 7 (cont'd)

[1]	[2]	[3]	[4]	[5]	[6]
715	386	8.4	149	9.8	17.3
769	415	8.4	160	10.1	18.3
832	449	8.4	177	10.6	19.3
891	481	8.4	192	11.1	20.3
954	515	8.4	205	11.5	21.3
1027	555	8.4	220	11.9	22.3
1085	583	8.4	238	12.3	23.3
1154	528	8.4	244	12.5	24.3
1230	664	8.4	250	12.6	25.3

Wind Speed 30 meters/second

100	54	3.6	55	6.0	4.0
120	65	4.0	62	6.3	4.5
140	76	4.3	68	6.6	5.0
160	86	4.6	74	6.9	5.5
180	97	5.0	78	7.1	5.9
200	108	5.3	86	7.4	6.4
220	120	5.6	91	7.7	6.6
240	130	5.9	97	7.9	7.2
260	140	6.1	101	8.1	7.6
280	151	6.3	106	8.2	8.0
300	152	6.5	111	8.4	8.4
400	215	7.6	133	9.2	10.2

TABLE 7 (cont'd)

[1]	[2]	[3]	[4]	[5]	[6]
500	270	8.6	154	9.9	11.9
600	324	9.6	173	10.5	13.4
700	378	10.2	190	11.0	14.9
800	432	11.1	208	11.5	16.3
900	486	11.8	224	12.0	17.7
1000	540	12.5	239	12.4	19.0
1100	594	13.1	255	12.8	20.3
1200	548	13.8	259	13.0	21.5
1300	702	14.4	283	13.4	22.7
1400	756	15.0	297	13.7	23.9
1500	810	15.5	310	14.1	25.0
1600	864	16.1	323	14.3	26.2
1676	905	16.8	336	14.7	27.0
1757	949	16.8	362	15.0	28.0
1832	1000	16.8	397	16.0	29.0
1940	1048	16.8	433	16.7	30.0
2035	1099	16.8	456	17.1	31.0
2144	1158	16.8	495	17.8	32.0
2232	1205	16.8	542	18.5	33.0
2335	1251	16.8	549	18.8	34.0
2448	1322	16.8	564	19.0	35.0

for a given run), and the run is sufficiently large, then the selection from the table is made according to the duration.

The use of the tables can be best clarified by concrete examples.

Example 1,

$V = 20$ meters per second; $D_v = 300$ kilometers; the duration of wind action is 15 hours.

First of all we look for a table in which the data for all the wave elements are computed for a wind speed of 20 meters per second. In Table 7 in the column "wind run" we find the number 300. On the line on which this figure is found and in the column "duration of wave growth" (identified in this case by the duration of wind action) we have the figure 9.5 hours. Comparing this figure with the given duration of wind action (15 hours), we may make the conclusion that in this case the development of waves is limited by the wind run, that is, already in 9.5 hours the waves will reach the greatest development in view of limitation D_v , inasmuch as, despite the 15 hour duration of wind action, the disturbance cannot exceed the values shown on this line in view of the limited run.

Selecting from this line the wave elements, we get: $h = 5.2$ meters; $L = 85$ meters; $T = 7.4$ seconds.

Example 2,

$V = 30$ meters per second; ~~$D_v = 300$~~ meters per second; the duration of wind action is 6 hours.

In the table computed for 30 meters per second, we find for

$D_v = 300$ kilometers the wind action duration is 8.4 hours, that is, greater than the given one (or the duration of wave growth), equal to 6 hours. In this case the development of a disturbance is limited not by the wave run, but by the duration of wind action, that is, the disturbance could develop even after 6 hours to 8.4 hours; therefore, the wave elements are taken from the line on which the duration of wind action is closest to 6 hours, that is, 5.9 hours.

From this line we take the values $h = 5.0$; $L = 78$; $T = 7.1$.

If it were necessary we could have taken h , L and T from the table, not for 5.9 hours, but, for exactly 6 hours by linear interpolation.

These examples are sufficient to understand the principle of constructing tables and using them for the conditions which were used as the basis of their computation.

Of course, the number of problems which can be objectively solved by their use can be considerably increased. For example, we can easily solve such problems as: what should the wind duration of speed V be for the wave height to reach a value h for a wind run D_v ; what should the wind run be so that the disturbance will reach height h , length L and period T for a wind speed V , and so forth.

With the aid of the tables reverse problems can be solved, for example, at given wave elements and a wind run to determine the wind speed, the duration of its action.

This circumstance has very great significance since it allows us to use the tables for computing wave elements even when we are

required to compute wave elements at a speed, duration of action and wave run when the wind begins to act not over a calm but over an agitated sea surface.

An application of the tables, unfortunately, is possible only when the direction of the disturbance acting previously does not deviate more than 35 degrees from the direction of a newly arisen wind.

In the latter case we can enter the new given wind speed on the table and determine what wind duration answers the wave height observed at sea at the start of the action of the given wind speed. After the wind action duration corresponding to the observed disturbance is determined we can compute the wave elements which will form under the wind action of a given speed in the usual manner, ~~but~~ counting the given duration of wind action not from zero but from that time which corresponds to the disturbance observed at sea.

Example 3.

Let $V = 20$ meters per second; $D_p = 300$ kilometers for a duration of wind action of 5 hours for a speed of 20 meters per second. However, up to the time when the wind will begin acting at sea with a speed of 20 meters per second, we note a wave with a height of 1.5 meters spreading in the same direction in which the wind of a speed of 20 meters per second will blow.

From Table 7, computed for a wind speed of 20 meters per second, we determine that to create waves with a height of 1.5 meters about two hours would be required. The given duration of wind action is 5 hours, therefore, since the wind would already have acted for

two hours, the general result will correspond to the wind action for $5 + 2 = 7$ hours. The disturbance for this would have the elements; $h = 4.2$ meters; $L = 65$ meters; $T = 6.5$ seconds.

Such wave elements require a run of not less than 200 kilometers, but according to convention the run equals 300 kilometers, that is, the run does not limit the wave development.

In this manner we establish that in 5 hours of wind action a disturbance will take place having elements corresponding to the number of hours of wind action duration.

For such a type it is necessary to have computation always in the form of a wave run. If in the given example the wave run were not 300 kilometers, but 120, say, then in this case the disturbance would have a height of 3.2 meters. It could not reach a greater height in view of the limitation of the run. And the height of 3.2 meters would be reached not in 5.1 hours but in 3.1 hours.

Table 7 applies in computing wind waves. The computation for elements of a swell should be made from a different table of L. F. Titov's. Part of it is shown in Table 8. [Table 8 on following pages]

The data we enter on this table is the time during which a swell spreads, L_0 and T_0 , that is, the starting wave length and period. The waves in spreading further ~~decrease in height, but~~ their period and length grow. In the table we have length L , period T , swell run D_3 and the ratio of heights $\frac{h}{h_0}$, that is, of the height at a given point of time h to the starting height h_0 .

Let us examine example 4.

TABLE 8 (Left hand part)

The time of
wave spread
of a swell,
hours

$L_0 = 70$ meters, $T_0 = 6.7$ seconds

$L_0 = 90$ meters, $T_0 = 7.5$ seconds

	L	$\frac{h}{h_0}$	T	D_3		L	$\frac{h}{h_0}$	T	D_3
[1]	[2]	[3]	[4]	[5]	[6]	[7]	[8]	[9]	
1	76	0.65	7.0	40	97	0.68	7.9	40	
3	90	0.59	7.6	120	116	0.62	8.5	140	
5	104	0.53	8.1	220	132	0.57	9.2	250	
7	114	0.48	8.5	300	148	0.53	9.7	340	
9	123	0.45	8.8	400	160	0.49	10.1	450	
11	132	0.41	9.2	510	170	0.45	10.4	570	
13	138	0.38	9.4	610	178	0.42	10.7	590	
15	146	0.35	9.7	720	187	0.39	11.0	820	

TABLE 8 (cont'd, right hand part)

L ₀ = 100 meters, T ₀ = 8.0 seconds				L ₀ = 110 meters, T ₀ = 8.4 seconds				L ₀ = 120 meters, T ₀ = 8.8 seconds			
L	$\frac{h}{h_0}$	T	D ₃	L	$\frac{h}{h_0}$	T	D ₃	L	$\frac{h}{h_0}$	T	D ₃
[10]	[11]	[12]	[13]	[14]	[15]	[16]	[17]	[18]	[19]	[20]	[21]
108	0.70	8.3	40	120	0.71	8.8	50	127	0.72	9.0	50
127	0.63	9.0	150	143	0.65	9.5	160	153	0.67	9.9	170
148	0.59	9.7	260	163	0.60	10.2	270	175	0.62	10.6	280
153	0.55	10.2	360	181	0.56	10.8	380	192	0.58	11.1	400
176	0.51	10.6	480	195	0.52	11.2	510	209	0.54	11.6	520
190	0.47	11.0	600	209	0.49	11.6	530	223	0.51	12.0	650
198	0.44	11.3	730	221	0.45	11.9	770	235	0.47	12.3	790
206	0.41	11.5	860	230	0.43	12.2	910	245	0.47	12.5	240

Example 4.

A wind of 30 meters per second force had a run of 500 kilometers and continued for 10 hours. We shall determine the elements of a swell at the end of the run at 3 and 5 hours after the wind subsides.

Using Table 7, we determine what the elements of the wind wave were at the end of the run at 10 hours after the wind started. From the table we get that in 10 hours the wave will have the following elements: $h = 7.6$ meters; $L = 133$ meters; $T = 9.2$ seconds. On the same line the run is 400 kilometers, but we are given 500 kilometers; consequently, in this case the limiting factor is the duration of wind action and not the run.

It is necessary to clarify what dimensions the waves which will come to the wind's final point will have 3 hours after the wind, acting simultaneously over the entire run, subsides, that is L_0 and T_0 . This problem must be solved by the following approach. It is quite obvious that at a 3 hour travel from the final point of a run at the time the wind subsides the waves will have a smaller length than at the end of a run. At the end of a run the waves will die down, but ⁱⁿ 3 hours waves will appear which at the time the wind subsided were at a distance of their 3-hour travel. The length of these waves are to be determined.

For this a wave length corresponding to a duration of wind action of 3 hours less than the given one is selected from Table 7, that is, $10 - 3 = 7$ hours. In Table 7 for a duration of wave growth of 7.2 hours we find the wave height to be 57 meters. This wave will

Example 4.

A wind of 30 meters per second force had a run of 500 kilometers and continued for 10 hours. We shall determine the elements of a swell at the end of the run at 3 and 5 hours after the wind subsides.

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It is necessary to clarify what dimensions the waves which will come to the wind's final point will have 3 hours after the wind, acting simultaneously over the entire run, subsides, that is L_0 and T_0 . This problem must be solved by the following approach. It is quite obvious that at a 3 hour travel from the final point of a run at the time the wind subsides the waves will have a smaller length than at the end of a run. At the end of a run the waves will die down, but ⁱⁿ 3 hours waves will appear which at the time the wind subsided were at a distance of their 3-hour travel. The length of these waves are to be determined.

For this a wave length corresponding to a duration of wind action of 3 hours less than the given one is selected from Table 7, that is, $10 - 3 = 7$ hours. In Table 7 for a duration of wave growth of 7.2 hours we find the wave height to be 57 meters. This wave will

spread as a swell, that is, its length and period will gradually grow, the height will decrease, and consequently, it will spread at a higher speed than the second wave. Furthermore, it becomes clear what distance a swell with the given starting length will travel in 3 hours. Now let us look at the table of swells calculations (Table 8) and we find data corresponding to $L_0 = 97$ meters ≈ 100 meters. In column D_3 we determine what distance a wave with a starting length of 100 meters will travel in 3 hours. This distance is 150 kilometers. In the first computation the wave run was not computed, but only the wind action duration was taken into account. On the line on which the duration of wave growth of 7.2 hours is shown, in Table 7 in column "wind-run" 240 kilometers is shown, but in Table 8 we again refer to Table 7 and determine the wave length for the run $500 - 150 = 350$ kilometers, by interpolation we find $L = 122$ meters.

Again we refer to the table with this wave length to compute the swell (Table 8) and we find $L_0 = 120$ meters for a 3-hour spread, $D_3 = 170$ kilometers. Now from Table 7 for a run of $500 - 170 = 330 \approx 300$ kilometers we can select all the necessary elements of a starting wave $L_0 = 111$ meters, $h_0 = 6.5$ meters, $T_0 = 8.4$ seconds.

From the table for the computation of a swell we will find conclusively that 3 hours after a wind subsides at the end of a 500 kilometer run we observe a swell with the elements: $L = 143$ meters; $\bar{h} = 0.65$ meters, $h = 6.5 \times 0.65 = 4.2$ meters; $T = 9.5$ seconds. It is clear that 3 hours after a wind action stops at the end of a 500 kilometer run the following changes will take place. Instead of a wind wave we will observe the correct swell. The height will de-

crease from 7.6 to 4.2 meters, the wave length will grow from 133 to 143 meters, the period will likewise increase from 9.2 to 9.5 seconds.

In a similar manner we can determine which swell will be observed at the end of the run in 5 hours. To do this, first of all we determine the difference $10 - 5 = 5$ hours and from Table 7 we find $L_0 = 68$ meters. From Table 8 for $L_0 = 70$ meters and $t = 5$ hours we get $D_3 = 220$ meters. Again in Table 7 for a run $500 - 220 = 280$ meters we find $L_0 = 106$ meters, $h_0 = 6.3$ meters, $T_0 = 8.2$ seconds.

In Table 8 there is no $L_0 = 106$ meters and $T_0 = 8.2$ seconds. But if we make $L = 106 \approx 110$ meters then in this case for $t = 5$ hours, $D_3 = 270$ kilometers. Returning to Table 7 for a run of $500 - 270 = 230$ kilometers we would find conclusive starting values of $h_0 = 5.8$ meters, $L_0 = 94$ meters, $T_0 = 7.8$ seconds. For these starting values in Table 8 we find by interpolation between the figures for $L_0 = 90$ meters and $L_0 = 100$ meters that $L_0 = 140$ meters, $\frac{h}{h_0} = 0.58$, $h = 3.4$ meters, $T = 9.5$ seconds.

22. CONSTRUCTING EQUATIONS FOR FORECASTING A DISTURBANCE IN CASE OF A LIMITED RUN

The relationships examined above are true only for those cases when the sea depth is equal to or greater than the wave length. When the sea depth is less than half the wave length the above mentioned theoretical relationship between the wave elements are not exact. Waves in shallow areas become shorter and steeper, the wave period and speed also changes. Therefore, for shallow basins having, as they frequently do, short runs, we recommend that empirical equations based on the products of the observations made in the same

basin be constructed. Empirical equations based on the products of observations must be constructed for off-shore areas as well.

In constructing such equations a careful preparation of the products of observations is made first of all.

The wind run can be determined by the isobars which are drawn on synoptic or hydrosynoptic charts. The length of a run corresponds to that section of an isobar which can be considered to be a straight line. For a greater definition, the section of an isobar whose tangents to any two points of the isobar will cross so that two of the four angles formed by the crossing of the tangents are less than 35 degrees, should be considered as a straight line section.

If the equations are worked out for a sea, bay or strait in which the distance from shore to shore and consequently the maximum wind runs are something like 100 kilometers, then we can avoid referring to synoptic and hydrosynoptic charts for determining a run, but accept the distance to the shore as the figure for a run in a direction opposite to the air current creating the disturbance. So, for example, N. D. Shishov in constructing equations for computing wave elements depending on wind speed and direction for the Gulf of Nevskiy used the wind run rose shown on Figure 30.

In case of a limited run all the observations of wave elements are grouped according to the directions of the wind run rose, the directions of the rays of which coincide with the wind directions at the point where the observations of wave elements were made.

After such a layout only those cases where the wind had char-

acteristics of a steady wind are carefully selected from the products of observations, that is, the wind neither increased nor decreased and did not change its direction for the entire length of the given run. Only this data of wave elements is considered as corresponding to a steady wind force and direction. The products of observation selected in this manner serve in constructing equations of the type:

$$h = f_1(V); L = f_2(V); C = f_3(V); T = f_4(V)$$

for each wind direction.

N. D. Shishov obtained the following equations for computing waves for several wind directions:

$$\begin{aligned} W - NNW \dots h &= 0.068V - 0.03, \\ NNW - NNE \dots h &= 0.058V - 0.02, \\ NNE - ESE \dots h &= 0.051V - 0.02, \\ S - WSW \dots h &= 0.048V - 0.02. \end{aligned}$$

Here h is the mean wave height in meters; V is the wind speed in meters per second.

By a mean height we mean the wave height obtained by dividing the sum of 10 successive well defined waves by 10. N. D. Shishov gives the changeover from a mean wave to a maximum in the form $h_{\max} = 1.35 h_{\text{mean}}$.

It is easy to see that in the above indicated relationship of wave height to wind speed the multiplier of V changes with the change in wind direction. The changes in this value, as observations have shown, occur as a result of changes in wind run.

This equation has the following form:

$$k = 0.025V\sqrt[3]{D},$$

where D is the length of a run in kilometers.

The general equation of wave height for all wind directions in the area of the northern dam head of the sealane in the Gulf of Nevskiy was:

$$h_{\text{mean}} = 0.025V\sqrt[3]{D},$$

In a similar way equations for other wave elements were obtained from the wind speed for this area.

The examined relationship are good only for the region where observations were made. In the Gulf of Nevskiy depths do not exceed 6 - 5 meters and the length of a wind run is less than 20 kilometers. N. D. Shishov found equations for a wind run of about 100 kilometers for sea depths varying up to 50 and more meters by the method described above. These equations have the following form:

$$h = aV\sqrt[3]{D}$$

$$L = bV\sqrt[3]{D}$$

for which

$$a = 0.0151 \cdot H^{0.342},$$

$$b = 0.104 \cdot H^{0.573}.$$

The relation of the values a and b to the mean sea depth in meters is shown in Figure 31. In examining these figures, the coefficients a and b change only up to a depth of 30 meters and for a further increase in depth remain practically unchanged. However, it is necessary to remember that these equations are obtained em-

pirically for runs not exceeding 100 kilometers. Using these relationships for large wind runs is probably impractical.

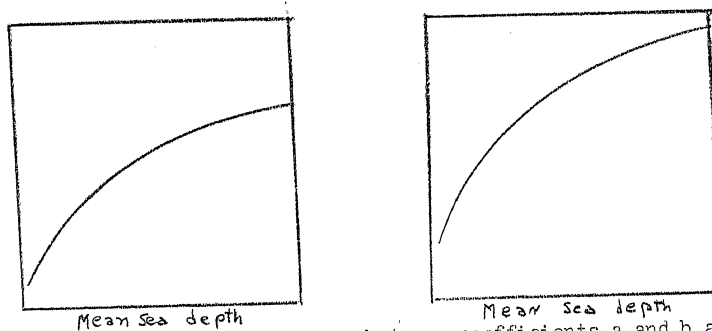


Figure 31. The relationship between coefficients a and b and the mean sea depth.

To determine the duration of wave growth or the time during which a wind must act, not changing its direction or speed, for a disturbance to reach a maximum value corresponding to a given wind speed. N. D. Shishov offers the following empirical formula:

$$t = \frac{0.33 \sqrt[6]{D^5}}{n \sqrt{bV}}$$

In this relationship t is the time in hours; D is the run in kilometers; V is the wind speed in meters per second; $b = 0.104H^{0.573}$; n is taken from the table shown below in a relationship between the mean and given run of sea depth (H) in meters: ~~[see table on following page]~~

[see below]

In this manner the duration of wave growth increases with an increase in the run and decreases with an increase in sea depth and wind speed.

H	-	2-5	5-10	10-15	15-20	20-30	30
n	-	0.95	0.98	1.02	1.05	1.09	1.12

Computing the direction of a disturbance does not present any fundamental difficulties thanks to the studies of V. V. Shuleykin. When it is necessary to compute disturbances in an open sea we can always accept that a disturbance spreads exactly in the direction in which the wind is blowing. In computing the direction from the shore we must use V. V. Shuleykin's formula allowing us to compute the direction of waves at the left end of the off-shore area. This formula has the following form:

$$\sin \alpha_0 = \frac{1 + \frac{mT^2}{H_1}}{1 + \frac{mT^2}{H_0}} \sin \alpha_1.$$

Here H_1 is the depth at point S, and H_0 at point A; T is the wave period at point S; $m = 0.05$; α_1 and α_0 are the angles which form lines between them drawn perpendicular to the wave crest and the shore line at point S and A (Figure 32). With this formula it is easy to find the direction of a wave at a given point in an off-shore area, if we know the direction of a wave in the open sea corresponding to α_1 , wave period T and depth H_0 at the point for which the direction of a wave is being computed and in the open sea depth H_1 .

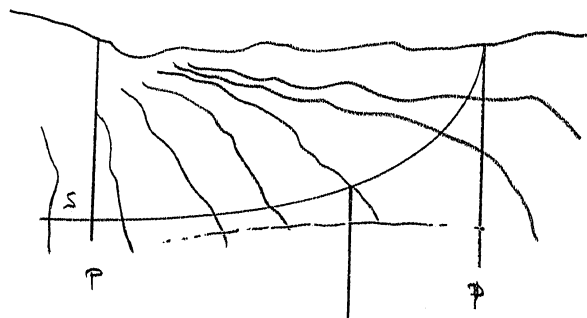


Figure 32. The change in direction of waves when approaching a shore

If the depth of an open sea is greater than the wave length, then H_1 may be considered infinitely large; in this case the second component in the numerator of the right side will become zero. Then the above written formula can be changed to a simpler form

$$\sin \alpha_0 = \frac{H_0 \sin \alpha_1}{H_0 + m T^2}$$

If $\alpha = 90$ degrees, that is, the disturbance spreads parallel to the shoreline, then the formula assumes a still more simple form:

$$\sin \alpha_0 = \frac{H_0}{H_0 + m T^2}$$

Using these relationships we can compute the angles at which the waves of an open sea approach the shore. But we should bear in mind that the wind acting along the shore causes its own wave system which interferes with the waves coming from the open sea. The wave system formed directly at the shore has considerably smaller heights.

24. CONSTRUCTING THE CHARTS FOR FORECASTING A DISTURBANCE

It is self-evident that disturbance forecasts can be made only on the basis of a forecast of a wind or a barometric condition. All the above mentioned equations were founded on the assumption that the wind and the disturbance are steady, that is, they do not change with the lapse of time. Actually, though, wind and disturbance are very changeable in time and distance. This is one of the difficulties we will encounter in developing disturbance forecasts even when equations are given for wind and disturbance.

A second difficulty in developing disturbance forecasts is the limited possibilities which synoptics uses in developing wind forecasts. In the general synoptic wind forecasts for 24 hours the following may be seen frequently: northerly wind 4 - 5 balls, increasing towards the end of the period to 7 balls. This forecast refers to some area of the sea defined on a chart by a rectangular geometric figure. For the adjacent area likewise defined by several straight lines a forecast may indicate; wind, northeastern quadrant, 6 - 7 balls, decreasing towards the end of the period to 4 - 5 balls.

Of course it is impossible to use such wind forecasts in developing forecasts of wave elements as the wave elements are computed by comparatively exact quantitative relationships which must use exact and definite values of length of a run (which is not indicated at all in the general synoptic forecast), wind speeds and duration. It is also necessary to know the changes in the areas of wind speeds. According to a synoptic forecast, thanks to the zoning, the wind appears to change in jumps from area to area even though this is not the actual situation.

In regard to the stated peculiarities of synoptic forecasts a specialist must determine the predicted wind field himself by developing disturbance forecasts from the data of a synoptic forecast and by consulting the synoptician.

The materials used in computing a wind are: a chart of the future barometric condition on the basis of which the weather forecast is written, and a synoptic chart of the actual condition for the last period of observation. To compute the wind on a blank large scaled chart including the entire sea with the adjacent shores, we draw lines of longitude and latitude. How thickly we draw them in depends on the size of the sea and the scale of the map. The network

of latitude and longitudes is determined once for a long period of time. This network is taken from the blank chart of a smaller scale and from this it is copied onto tracing paper or wax paper. For the sake of brevity the parallels and the meridians are assigned consecutive numbers. Numbering the parallels starts from the north, numbering the meridians from the west.

Further, from the synoptic chart (good for the last period and the coming one) we take the pressure for the points where the parallels and the meridians cross. To do this we use the network of meridians and parallels traced on the tracing paper. Inasmuch as the isobars oftentimes do not coincide with the intersections of the meridians and parallels we use the general linear interpolation of pressures to determine the pressure at these points.

The pressure taken in this manner for each point of intersection of a parallel and a meridian is transferred to millimeter paper of corresponding scale. In doing this the pressure taken from the actual synoptic chart and the pressure taken from a chart of the future condition are transferred to graphs so that they are separated by 24, 30 or 36 hours, depending on the number of hours ahead for which the chart of future barometric condition is being made up. The points on the graph corresponding to the actual and future pressure are connected by straight lines. From these graphs we take the pressure for intermediate periods following one another by 6 hours (we need not apply graphic interpolation, for we can solve the problem mathematically).

As a result of these operations a table is compiled in which we have several values of atmospheric pressure indicated for each

point of intersection of a meridian and parallel for 6-hour periods. The next operation is computing mean pressure values. So, if in the tables we have the pressures for 3, 9, 15, 21, 3, 9 hours then a mean pressure is computed for each point of intersection for periods of 3 and 9, 9 and 15, 15 and 21, 21 and 3, 3 and 9 hours. When the pressure changes slowly during the time for which it is predicted, and this occurs quite frequently, we can limit ourselves to one or two mean values. For this we take the extreme values, for example for 3 and 9 hours of the following 24 hours. In these cases it is less necessary to construct graphs or mathematical interpolations. In this manner, the necessity of interpolating and the number of periods for which averages will be made must be determined each time depending on the complexity of the barometric condition and on its change for the period for which the disturbance forecast is to be made. In an overwhelming number of cases we can ignore the interpolation. Using the data of average pressure values the projections of wind currents are computed. The wind current is projected on to parallels and meridians. The computations are made by the formula

$$V = \frac{a}{\sin \phi} \frac{\Delta p}{\Delta S}.$$

Here V is the wind speed; ϕ is the latitude of the locale Δp is the pressure differential of two adjacent points of intersections of meridians and parallels; ΔS is the distance between these points; a is a constant coefficient.

Inasmuch as the angles of wind deviation from an isobar over a sea are not large, about 10-20 degrees, and not exactly steady at these values either, it is well to assume that the wind over a sea

always blows along an isobar.

It is most desirable to make special computations of the coefficient for each area of the sea where we may expect distortions in the wind pattern. Areas where we may expect distortions are bays, straits, off-shore high mountain regions etc. When we undertake the study of changes in the coefficient a for separate areas of a sea in many cases the above-mentioned relationship for computing a gradient wind must be completed by one more component, namely:

$$V = \frac{a}{\sin \phi} \frac{\Delta P}{\Delta s} + b \frac{\Delta T}{\Delta s}$$

The component $b \frac{\Delta T}{\Delta s}$ should be considered when off-shore areas are being examined. The temperature differential between the water and air is denoted by $\frac{\Delta T}{\Delta s}$. Including this differential we can estimate approximately the effect which V. V. Shuleykin theoretically studied. Here we are speaking of the fact that the wind is always stronger at sharp capes projecting well out into a sea and weaker in bays. This phenomenon is well known. We need only recall the frequently raging storms at Cape Horn, at Cape Igol'niy, etc.

It should not be thought that the effect of increased winds is noticed only when the ends of continents are studied which project deep into an ocean. Even in cases of drawn out islands or peninsulas this phenomenon is clearly noted. Such are, for example, Cape Zhelaniya on Novaya Zemlya or the Apsheron Peninsula in the Caspian Sea on which Baku is located. Navigators who quite frequently experience difficulties with storm winds know well that when navigating on an open sea in calm weather they can frequently encounter 7-8 and even 9 ball winds on the approaches to the Apsheron Peninsula.

The pressure differential computed by the first component of the above written relationship is determined by the column of air stretching from sea level to the upper limit of the atmosphere. Of course this differential cannot fully reflect the phenomenon which the lower layers of air experience in their sharp transfer from one lower level to another (from land to sea or from sea to land). This effect cannot be said to be for the entire thickness of atmosphere just as it cannot be said that the daily movement of air temperature over sea and land creates a breeze phenomena. Obviously when computing breezes it is also necessary to use the second component of the relationship under discussion.

Suppose that a forecaster has at his disposal all the data for computing wind. In this case it is necessary to consider the biggest possible simplification of the computations connected with forecasting disturbances and wind. Above all this can be achieved after having made all the calculations before making a pressure forecast.

After completing the computations of the wind projection on parallels and meridians or simultaneously with the completion of these computations on a blank chart of a sea between a corresponding pair of points, the values of wind speed are expressed in figures, but the direction of the wind projections is indicated by arrows. The projection should use this principle : ~~a wind blows in the direction in which an observer is facing if he stands so that the high pressure is on his right and the low on his left.~~

Having finished placing the speed and direction projections of a wind on a chart we construct the lines of flow. To do this we

break up the distance between the points of intersection of parallels and meridians into a number of parts proportional to the projections of wind speed.

Suppose that a projection of wind speed is equal to 5 meters per second then the distance between the points is divided by lines into 5 even parts. If it is 10 meters per second the breakup is made in 10 equal parts. Fractional values of wind speed are rounded out to the nearest whole number.

The final stage in constructing a chart of lines of flow is drawing smooth lines through the divisions made proportional to the wind speeds. A line must be drawn through each division. Having drawn the smooth lines which will be the lines of flow, arrows are used to indicate the direction of the wind speed projections.

Because the vector representing the wind must always be directed along the flow next to the line, but the movement is considered two-dimensional, and bearing in mind that the wind speeds are unevenly dispersed over an area, the lines of flow will break off in some areas and in other begin anew.

The charts of lines of flow obtained in the manner described (they are constructed for all those periods for which average pressure values are computed) fully characterize the predicted field of wind speeds.

The direction of an air flow is followed through by the directions of the lines of flow, and its speed by their thickness.

For simplicity of analysis, before constructing a chart of

lines of flow, information on ice conditions is transferred to the blanks on which the lines of flow are made (it is very important to know the edge and the thickness of ice), as well as information on disturbances, level and temperature of water and air, determined for the last period of observation.

25. WORKING OUT DISTURBANCE FORECASTS

The charts of lines of flow are very convenient for the computations which must be made in forecasting disturbances. All the elements of a wind flow (speed, direction and run) are represented on these charts, and a series of charts gives a representation of the change in a wind flow in time and expanse.

The construction of these charts requires a considerable effort and, what is especially important, this effort must be made in the shortest possible time as there is no sense in doing it any other way. That is why the work of constructing a chart of lines of flow must be carefully thought out and simplified as much as possible. All possible calculations must be made in time, blank charts carrying the corresponding data must be prepared, as must the forms for the calculations and so forth.

Data is taken from the latest synoptic chart by the forecaster immediately after its analysis. The most intensive work begins after the chart of the future barometric field is constructed by the forecaster. The entire work from this moment on is conducted by a specialist with the aid of one or two technicians. The work can be completed in a short time only when the workers have sufficient experience in the entire cycle of the work.

Above we had the formulas relative to wave elements with winds for off-shore shallow areas with limited runs. Similar relationships can be constructed for other sea areas if we have the necessary observations required in constructing these relationships. ^{For} ~~By~~ avoiding rather complicated computation, tables are computed for each forecast in advance for the corresponding formulas. These tables contain -- depending on the wind speed, wave run, sea depth and the continuance of wind action -- the wave elements: height, length, period, speed and steepness. Such tables compiled by L. F. Titov have been obtained for deep-water seas with long runs.

CHAPTER VI

FORECASTING WATER TEMPERATURE AND ICE FORMATIONS

26. COMPUTING THE ELEMENTS OF THERMAL EQUILIBRIUM OF A SEA SURFACE

The variations in water temperature, the changing of water from a liquid to a solid, that is, the formation of ice on a sea surface, the melting of an ice cap are the reasons for the inconsistency of a thermal equilibrium. Therefore, in order to understand thermal processes occurring at sea it is necessary to understand first of all which elements make up a thermal equilibrium.

First, V. V. Shuleykin showed in the example of the Kara Sea how a computation may be made of the elements of a thermal equilibrium. Computing the thermal equilibrium of the Kara Sea he concluded that a stream of warm current had to enter the Kara Sea carrying a definite amount of heat into it. In the absence of this warm current the thermal equilibrium of the Kara Sea would be negative for many years, after which there would be an unavoidable freezing of the Kara Sea to the very bottom. As a result of observations by an expedition a branch of the warm Atlantic current was actually found entering the Kara Sea from the north. In this manner the theoretical reasoning of V. V. Shuleykin and his computations were fully confirmed.

A thermal equilibrium of a separate section or of the entire sea consists of the following elements:

1. The receipt of thermal energy in the form of direct and

diffused solar radiation.

2. The loss of heat in the effective radiation from a sea surface.

3. The loss of heat by evaporation.

4. The receipt of heat as a result of the condensation of water vapor on the water surface.

5. The heat exchange with the air -- positive (the sea is warmer than the air, the sea loses heat).

6. The heat exchange with the air -- negative (the sea is colder than the air, the sea receives heat).

7. The heat exchange within the water created by the horizontal travel of heat together with the water masses.

8. The heat exchange within the water created by convection and the turbulent mixing of water masses.

9. The receipt of heat as a result of ice formation.

10. The loss of heat as a result of ice melting.

The variations during all the enumerated elements of a thermal equilibrium can be studied directly by means of appropriate measuring instruments. However, most of these observations require complicated apparatuses, high qualifications of the observer and are conceived with a great many other difficulties. As a rule, the entire complex of observations at one place necessary in determining the path during the enumerated elements of a thermal equilibrium is not made, and the separate elements are observed comparatively in-

frequently. That is why, in practice, in an overwhelming majority of cases it becomes necessary to use not the observed values of the elements of a thermal equilibrium but to compute them.

The loss of heat or its influx depends on these or other hydro-meteorological conditions. So, for example evaporation and condensation depends on the humidity deficit and wind speed, heat exchange depends on the differences in water and air temperatures and wind speed and so forth.

Most of the elements of the thermal equilibrium can be computed by the usual hydrometeorological observations made by second class stations.

If we disregard the internal heat exchange, then in computing the thermal equilibrium we can use the following relationship:

$$Q_w = Q_0 - R_w \pm Q_k \pm Q_{T.O.},$$

where Q_0 is the receipt of direct and diffused solar radiation; R_w is the effective radiation from the water surface; Q_k is the evaporation or the condensation from the water surface; $Q_{T.O.}$ is the positive and negative heat exchange of the water surface with the air.

To compute the direct and diffused solar radiation we can use Table 9 compiled by V. N. Ukraintsev. In Table 9 we have the maximum values of the maximum daily totals of the full solar radiation for various latitudes in calories per day per square centimeter. [See Table 9 on following page]

TABLE 9

Latitude in degrees	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
35	424	505	620	750	840	(880)	845	780	650	540	445	395
40	325	415	550	680	770	(810)	780	700	575	445	350	290
45	210	320	460	610	715	760	720	633	500	355	250	190
50	115	220	390	560	690	740	690	590	434	285	160	100
55	72	170	330	510	665	720	660	550	380	230	120	65
60	50	130	286	475	645	700	640	512	334	180	80	35
65	2	85	240	446	637	690	615	460	280	130	30	-
70	-	58	200	445	647	710	610	420	235	95	-	-
75	-	-	180	456	710	800	690	430	190	88	-	-

To obtain the values of radiation for intermediate latitudes we must apply graphic interpolation. We must bear in mind that the amounts of radiation shown in the table refer to the average for a month. Therefore, to obtain the radiation values for each day we must also apply interpolation. The best results are obtained from graphic interpolation when a smooth curve of the path of radiation is traced out for the period of an entire year or at least for half a year, so that the interpolated part is within the limits of the traced curve.

In order to change from the maximum daily values of the sums of solar radiation to the actual solar radiation passing through a water surface we must correct the table values for the humidity effect and also consider the fact that part of the energy falling on water surface is reflected (water albedo). To do this we may use the following empirical relationship:

$$Q_0 = (\Sigma Q' + \Sigma q)_0 (1 - cN)(1 - r)$$

Here Q_0 is the receipt of the actual solar radiation passing through a water surface expressed in calories per day centimeters²; $(\Sigma Q' + \Sigma q)$ is the maximum daily total of a full solar radiation derived directly from the table indicated above or obtained by interpolating the data in the table; c is the coefficient dependent on the form of humidity, N is the amount of overcast in tenths of sky coverage, r is the albedo of the water.

The water albedo can be taken as an even 0.15 and when N designates the general cloudiness c should be taken as 0.70 or 0.75.

The effective radiation from a sea surface R can be computed

by the formula:

$$R_w = \alpha \sigma T_a^4 (1-A)(1-cN) + \alpha (\sigma T_w^4 - \sigma T_a^4)$$

In this empirical formula R_w is the effective radiation of the water in calories per square centimeter minutes; α is the radiation capability of the water (in the first approximation this value can be taken at an even 1); σ is the Stefan-Bol'tsman constant equal to 8.26×10^{-11} calories per square centimeters minute degree; $A = 0.071 \sqrt{e} + 0.53$, where e is the absolute humidity of the air expressed in millimeters; c is the coefficient dependent on the cloud form (it can be taken as equal to 0.75 if N is the amount of general cloudiness) N is the amount of overcast in tenths of sky coverage; T_w is the absolute water temperature in degrees; T_a is the absolute air temperature in degrees.

The computations for evaporation and heat exchange can be made by Sverdrup-Kuz'nin formulas. Even though these formulas have certain shortcomings in most cases they allow us to obtain values close to the actual ones.

The formulas for computing the amount of heat used in evaporation have the form:

$$Q_i = \frac{0.216 (E - e) V}{0.067 \log \frac{z_1 + z_0}{z_0} \log \frac{z_2 + z_0}{z_0} + \Delta z} \text{ calories/centimeters}^2 \text{ minute.}$$

The structure of the formula for computing heat exchange is similar to the formula for computing the heat loss in evaporation

$$Q_{T.O.} = \frac{0.034 (t_w - t_a) V}{0.067 \log \frac{z_1 + z_0}{z_0} \log \frac{z_2 + z_0}{z_0} + \Delta z} \text{ calories/centimeters}^2 \text{ minute.}$$

In these formulas the following designations were made: E is the maximum vapor tension in millimeters for the water surface temperature; e is the absolute humidity of the air in millimeters; V is the wind speed in meters per second; t_w is the water temperature in degrees; t_a is the air temperature in degrees; Z_1 is the height in centimeters at which the readings of wind speed are taken; Z_0 is the parameter of roughness in centimeters; ΔZ is the thickness of the laminar layer in centimeters.

The parameter Z_0 varies with the various physiogeographical conditions. P. P. Kuz'min determined the following values of the coefficient of roughness: for the deep-water section of the Kadaksha Bay of the White Sea $Z_0 = 1.55$ centimeters, and for the off-shore areas of the same bay $Z_0 = 0.5$ centimeters, for the area of Chechen' Island (Caspian Sea) where the sea depth varies 1-2 meters, $Z_0 = 0.15$ centimeters.

Ya. A Tyutnev established that for off-shore areas of the Azov Sea $Z_0 = 0.35$ centimeters.

For the thickness of a laminar layer ΔZ observers used the following values: $\Delta Z = 0.13$ centimeters (by Sverdrup in ocean condition), $\Delta Z = 0.20$ centimeters (by Davydov for Lake Sevan), $\Delta Z = 0.25$ centimeters (by Tyutnev for Taganrog Bay of the Azov Sea).

The formulas of Sverdrup and Kuz'min can be presented in a more simple form, for example, as it was done by Tyutnev for the conditions in Taganrog Bay:

$$Q_i = \frac{13(E-e)V}{0.95+0.25V} \text{ calories/cm}^2 \text{ day} \quad Q_{T.O.} = \frac{4.9(t_w - t_a)V}{0.73+0.25V} \text{ calories/cm}^2 \text{ day}$$

For this Z_1 is taken as equal to 350 centimeters and $Z^2 = 1500$ centimeters. Or, more simply, if we assume:

$$A_i = \frac{13V}{0.95+0.25V}; \quad A_{T.O.} = \frac{4.9V}{0.73+0.25V}$$

then in this case

$$Q_i = A_i(E-e); \quad Q_{T.O.} = A_{T.O.}(t_w - t_a)$$

The values of A_i and $A_{T.O.}$ may be presented in the form of the following table compiled by Tyutnev for the conditions at Taganrog Bay: (V units in meters per second).

V m/s	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
$A_{T.O.}$ "	5	8	10	12	13	13	14	14	15	16	16	16	17	17	18	18
A_i	10	18	23	27	30	32	34	36	37	38	39	40	40	41	41	42

The above table and formulas allows us to compute the thermal equilibrium of a water surface for any interval of time. We must remember that in the first approximation of the amount of heat expended in the condensation of water vapor on a water surface, we may use the formula serving computations of heat loss in evaporation. The experiments made to study quantitatively the

condensation processes showed that the errors in this case were within allowable limits.

27. FORECASTING WATER TEMPERATURES

In computing the thermal equilibrium of a water surface by the above formulas we can observe with sufficient accuracy the variations in water temperature. This is possible when we can disregard the horizontal interwater heat advection or in other words, the horizontal heat transfer and when we have no significant vertical stratification of water temperature.

Ya. A. Tyutnev showed in the example of the Azov Sea and Northern Caspian Sea the possibility of computing the changes in water temperature by computing the receipt and the expenditure of heat through a sea surface. In this, in the case studied by him for shallow sections of the Azov and Caspian Seas, the computations were made from observations at ^{the} Eysk, Osipenko, Kazantip and Astrakhan Roads stations. The computations included the period from April to October 1939. The changes in mean daily temperatures of a water surface from day to day were computed; the thermal equilibrium was also computed daily. The numerical values of changes in water temperature and of the daily sums of heat received or lost obtained in this manner were placed on graphs. In Figures 33 and 34 on the horizontal axis we have the daily sums of heat received and lost by a sea surface, and on the vertical axis we have the equilibriums of mean daily temperature values corresponding to these sums.

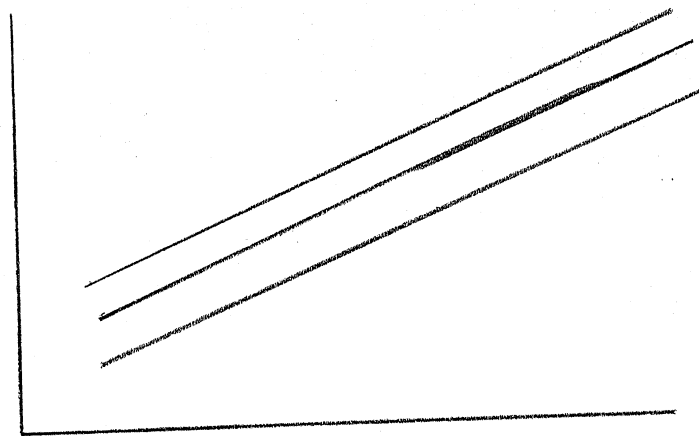


Figure 33. Relationship of the daily changes in water temperature to the thermal equilibrium of a sea surface made from the observations at Eysk Station.

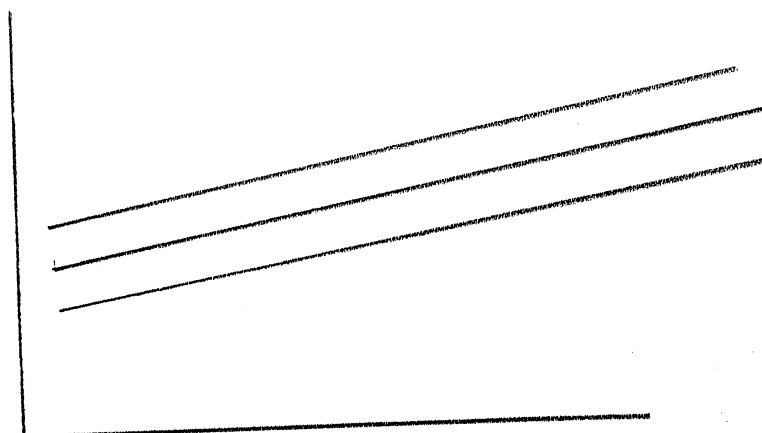


Figure 34. Relationship of the daily changes in water temperature to the thermal equilibrium of a sea surface made from the observations at Osipenko Station.

It is necessary to note that in making this type of a relationship for areas of various depths we must unavoidably find a different slope of the line $\Delta t_w = f(\Delta Q)$ to the abscissa, since the relationship is made between the amount of heat seeping through a water surface in a positive and a negative direction and the change in water temperature only at the sea surface. It is obvious that heat passing through a water surface of an area of 1 square centimeter is related, thanks to the turbulent thermal conductivity, not only to the change in temperature of the surface layer of water, but, in comparatively shallow areas, to the change in water temperature for the entire mass from surface to bottom. This is clearly observed in the mentioned relationships. The lines $\Delta t_w = f(\Delta Q)$ for Osipenko and Eysk pass through zero, but the change in water temperature of ± 1 degree for Eysk takes place when $\Delta Q = \pm 350$ calories, but at the same time for Osipenko the change in water temperature of ± 1 degree is dependent on receiving or losing 550 calories.

A comparison of the mean sea depths in the areas of the examined points reveals the physical reason for the noted difference in the number of calories involved in a change in the water temperature of 1 degree at the surface. The mean depth in the Eysk area is 350 centimeters and in the Osipenko area it is 550 centimeters. Since the thermal capacity of water can be taken as equal to one, this strange comparison, at first observation, of the number of calories with the number of centimeters of depth becomes understandable. This comparison indicates the good mixing of water in offshore areas, as a result of which, the heat gained or lost through 1 square centimeter of sea surface for each day changes the water temperature in its entire mass from surface to bottom. The data

on changes in water temperature for separate horizons obtained by the direct measurements at roadstead points fully confirm the conclusions made above.

Using the relationship $\Delta t_w = f(\Delta Q)$ we can make the computations for the changes in surface water temperature for several days. In the formulas indicated above for computing changes in water temperature for several days we must make successive computations from day to day. But roadstead departures are made not oftener than after 5 days, therefore, we cannot use the data of direct observations for each day. In this case, it becomes necessary to use the relationships $\Delta t_w = f(\Delta Q)$ made with the aid of daily off-shore observations to get the computed values of water temperature for each day. To compute the temperature for the first day of the five-day period we must use the water temperature of the last day of the previous five-day period. The thermal equilibrium for these 2 days allows us to compute the temperature change for the first day of the five-day period under examination. This change may be determined by a graph of the type in Figure 34. The water temperature of the second day of the five-day period is determined from the computed water temperature of the first day and the summary thermal equilibrium for the first and second day, and thus the computations are made to the last day of the five-day period. On Figure 35 we have the relationship between the changes in water surface temperature for a five-day period and the totals of thermal equilibrium for this same period. As is evident from the figure under consideration the relationship remains close even for five-day period temperature changes. The coefficients of correlation showing the relationship are equal to 0.85 and 0.88.

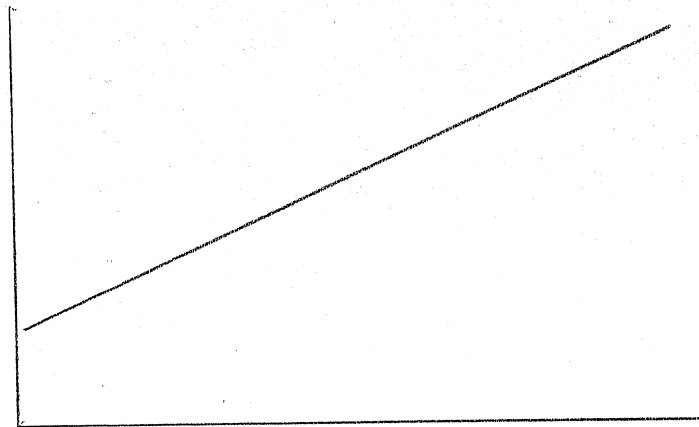


Figure 35. The relationship of the changes in water temperature for a five-day period to the thermal equilibrium of a sea surface made from observations at Astrakhan Roads.

It is necessary to note that in the relationship made for Astrakhan Roads the line of regression does not pass through zero (Figure 35). It appears that for this area the water temperature decreases for a small positive equilibrium. This decrease in temperature for a small positive equilibrium is explained by the fact that Astrakhan Roads is located in the sphere of influence of the colder waters of the Volga during the entire season. We can compute on the average approximately how much colder the mean Volga waters adjacent to the Astrakhan Roads are.

For the line of regression to pass through zero on the graph it must be moved parallel to itself about 360 calories to the left or be raised 0.7 degrees. If we bear in mind that the mean depth in this area is 500 centimeters and that in order to raise the water

temperature 1 degree for the entire thickness we must use 500 calories of heat for each one square centimeter of sea surface, then it will become apparent that on an average in the area of the roads we get waters 0.7 degrees colder than the surrounding waters and in heating them 360 calories are used. The value 0.7 degrees or 360 calories, of course, is not constant. It must vary with the time of year, the height of the level or water discharge and, finally with the weather conditions in the lower Volga and in the area of the roads including here also the influence of the sgonno-nagonnye winds. The "running-up" wind will help centrifuge, delay the inflow of Volga waters, and the "running-out", conversely, will accelerate the movement of the Volga waters to the roads.

All these changes in the flow of the Volga waters to the roads and the variations of their temperature may be computed with sufficient accuracy just as the heat or cold advection in other cases, when it is related to the well studied currents. Until the present time there are no well and thoroughly studied currents, hence, we cannot have a good enough example here allowing us to follow all the processes occurring at sea. However, some studies, though by indirect methods, clearly allows us to follow the influence of sgonno-nagonnye winds on the variation of water temperature.

As was indicated above, in a deep sea bordering on a shore line a wind blowing perpendicular to the shore does not change the level at the shore, because in this case the current arising at sea slides along the shore. Conversely, a wind blowing along the shore causes variations in sea level because the current arising from this

wind direction tends either towards the shore or away from the shore depending on whether the wind is blowing from left to right or from right to left relative to an observer facing seaward.

The speed of a subsurface current can be expressed by the simple relationship:

$$G = \frac{Bv}{\sqrt{\sin}} \cos \beta$$

Here G is the speed of the subsurface current; B is some constant coefficient; V is the wind speed; β is the angle between the wind direction and the shoreline ϕ is the latitude of the locality.

In this manner the speed of a subsurface current is directly proportional to the projection of the wind speed on the tangent to the shoreline. From here it follows that the amount of water carried by the subsurface current to the shore is directly proportional to the amount of air moving over the sea along the shoreline.

V. V. Shuleykin designed an apparatus which records separately the component normal to the shoreline, of the air flow into the tangent to the shoreline forming the air flow. A sample of a recording of this apparatus is borrowed from Shuleykin's, Fiziki morya (The Physics of the Sea) and shown in Figure 36. Here along the abscissa we have plotted the path travelled by air masses perpendicular to the shoreline (Q_n) and along the tangent to the shoreline (Q_t). The middle lines in both drawings separate, in the first case, the direction of the air flow from the sea (the recorded curve is above the middle line) and from the shore (the recorded curve is under the middle line), and in the second case, the "running-out" wind direc-

tions (the curve is under the middle line) and the "running-up" (the curve is above the middle line). In the same figure the lower curve shows a record of the water temperature at the surface of a sea having been drawn parallel to the recordings forming the wind flow.

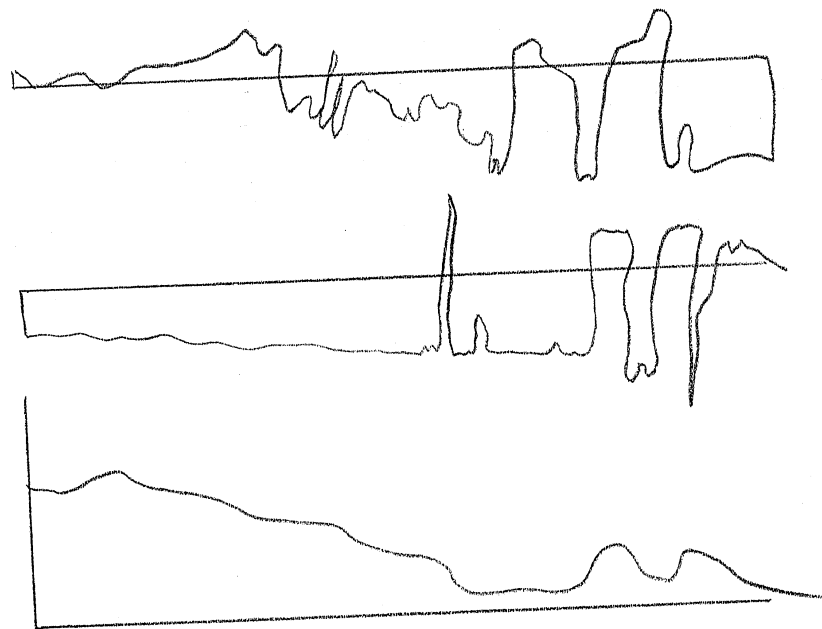


Figure 36. A recording of the path of air masses moving perpendicular to the shore (Q_n) and parallel to the shore (Q_t), and the changes in water temperature with the passage of time.

A careful examination of this figure shows the best comparison in the path of the curve of the surface water temperature and the curve registering the "running-out" and "running-up" of the water. Of course, when a "running-out" takes place the water temperature decreases because the warm surface waters move away from the shore

seaward, and in its place much colder water rises from the bottom. Conversely, the "running-up" wind direction brings about an increase in water temperature because the warm surface water coming from seaward builds up at the shore. These observations were made at the Black Sea Hydrophysical Station located on the Crimean shore at Katsivel'.

Ya. A. Tyutnev succeeded in constructing relationships based on the computation of the effect of "running-out" winds which can be successfully used in forecasting sharp changes in water temperature in Crimean coastal areas by wind speed and direction forecasts.

In the deep areas at the shores of Crimea the waters are marked by considerable vertical water temperature gradients during the summer. If we usually observe the water temperature at the sea surface as 22-24 degrees, then at a certain depth the temperature during the entire summer remains around 8-9 degrees. During "running-out" winds the warm surface water leaves the shore and in its place rise up waters with temperatures of 8-9 degrees. After the "running-out" wind stops the normal distribution of temperatures is quickly established. Using the products of usual observations by coastal hydrometeorological stations it was possible to construct for each area where the described phenomenon is observed, the relationships which allow us to predict sharp variations in water temperature. These relationships are similar to the relationships shown on Figure 37 where the sum of the projection of a "running-out" wind speed in a direction coinciding with the general shoreline direction is drawn on the abscissa, and the water temperature on the ordinate.

The sum of the wind projection is computed for each period of observation. Consequently, the longer the "running-out" wind acted and the greater its speed, the greater the sum will be.

It follows from Figure 37, that, depending on the growth of the wind speed sum, the water temperature at the surface drops sharply and after a certain value of the sum the temperature drop stops and for a further growth of the sum the water temperature remains unchanged. Apparently, this indicates that the subsurface water had risen to the surface of the sea.

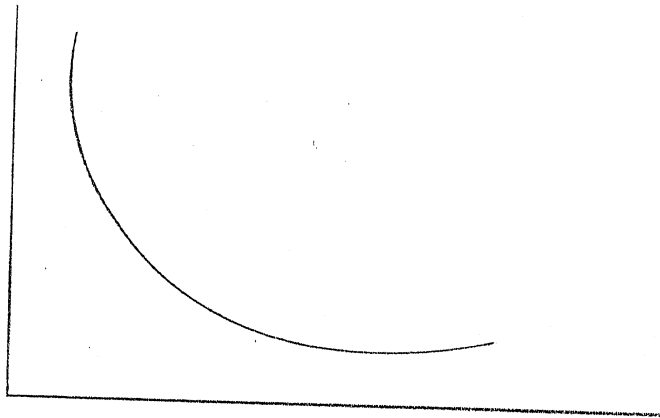


Figure 37. The relationship of the water temperature at the surface of the sea at a deep off-shore area to the "running-out" winds.
(units in meters per second)

Changes in water temperature at the surface of the sea during "running-out" winds occur so quickly that there is no necessity of using computations of thermal equilibrium in completing the relationship under examination. Relationships of this type may be used directly in forecasting sharp changes in water temperature. For this it is necessary to use only the wind forecast.

The relationship of variations in water temperature to variations in the thermal equilibrium of a sea surface allows us to make computations of temperature changes to a degree of accuracy sufficient in practice. However, it is still impossible to use the above relationships for computing the elements of a thermal equilibrium for forecasting water temperature. The trouble is that the air humidity in these relationships is not being predicted by forecasters, and that the wind and cloudiness forecasts, though they are being made, will have insufficient accuracy and dependability.

A more complicated matter is the forecasting of water temperature for those regions where it changes sharply with depth and area. Relationships similar to those above for the Crimean coast, probably, in many cases allow us to forecast the water temperature where there is a sharp temperature stratification. However, it can be said with great assurance that relationships of this type are not applicable in all cases. Obviously for stratified water, we cannot always assume that the heat passing through the sea surface changes the water temperature evenly for the entire depth, because during comparatively calm weather only some surface layer will be heated.

The question of just exactly which layer must be mixed, depending on the turbulent mixing related in turn to the current speed and wind speed, can not yet be decided in a general form. There could be times when a storm wind mixes the waters of a surface and subsurface layer which had not until this time been participating in thermal processes connected with the heat exchange through a sea surface. Clearly, the water temperature in this case will depend on the change in water temperature with the depth and on the depth which

the mixing will reach. All these problems have still only been very slightly worked out.

The calculation of the sgonno-nagonnnye wind effect, evidently, also cannot be as simple as in the case examined above. Probably, a careful study of the horizontal distribution of water temperature is frequently required. The solution of all these problems is being delayed because of the absence of the necessary observations.

Despite the difficulties of solving the problems in forecasting water temperature for sufficiently accurate weather forecasts for shallow sea areas it could be possible to begin a regularly planned servicing of interest organizations with forecasts of water temperature. However, forecasting elements of weather necessary for this are not yet made with sufficient accuracy and reliability for this purpose. All this taken together requires us to forecast water temperature not in a systematic pattern but only by special request for particular areas.

28/ FORECASTING THE APPEARANCE OF ICE

The matter of forecasting the appearance of ice, opening up and clearing the sea of ice, is somewhat simpler when compared to forecasting water temperature. Forecasting the appearance of ice is simplified thanks to the fact that during the period when the air and water temperatures are close to zero or below zero it is possible, with sufficient accuracy for practice purposes, to assume the thermal equilibrium of a water surface or an ice surface for several days to be proportional to the sums of air temperatures, and the latter is predicted better than any other weather element,

In addition, during periods close to the time of the appearance of ice and especially after the formation of ice on a sea surface we always observe a vertical homoisthermism and gomokhalinnost' [uniform salinity] in the large subsurface layer of a sea. The methods of short-term forecasting of characteristic ice phases developed in the Central Institute of Forecasting by P. P. Nikiforov, E. M. Sauskan, and Ya. A. Tyutnev, will be demonstrated below.

The method is based on empirical studies and is set forth on examples of separate seas and even of separate points but is sufficiently general and applicable for most of the sea areas of the Soviet Union, and, more than that, forecasting appearances of ice on rivers is carried out by a similar method. The method of forecasting ice on rivers was developed by V. D. Komarov and L. G. Shulyakovskiy. It is obvious that some modifications of a minor nature in the method are unavoidable and even necessary in computing, in each separate case, the peculiarities of physiogeographical conditions, the peculiarities of the thermal field of each area for which forecasts are being worked out.

First of all, we must clarify the physical concept of the relationship mentioned long ago between the periods of air temperature passing through zero degrees and the periods of the appearance of ice. These relationships in the practice of operational forecasting have been widely used for a long time, and only in recent years, thanks to the work of Komarov and Shulyakovskiy on short-term forecasting of the appearance of ice on rivers and the work of Nikiforov and Tyutnev on forecasting the appearance of ice on seas, have more accurate relationships between periods of ice appearances and sums of degree-days, of frost, have been introduced into practice.

Relationships of periods of air temperature passage through zero degrees to data of the appearance of ice, in a hidden form, are relationships of sums of degree-days of frost, computed from the time the air temperature passed through zero to the appearance of ice, to the periods of the appearance of ice, but only more coarse ones not considering the peculiarities of the variations in air temperature after its passage through zero and the heat reserve in the water at the time the air temperature passes through zero degrees. The consideration of the peculiarities of the passage of air temperature is coupled with the necessity of making ice forecasts from air temperature forecasts.

Practice in the operational work at the Central Institute of Forecasting has shown that, despite the comparatively low justification of air temperature forecasts, using these forecasts in developing forecasts of periods of ice appearance improves their quality in comparison with the forecasts from simple relationships of data to the actual passages of air temperature through zero degrees. It appears that the effect of considering the changing of air temperature after its passage through zero, as well as the heat reserve of water to this time is extremely essential in the sense of improving the quality of forecasts. Errors in ice forecasts decreased despite the inaccuracies in air temperature forecasts.

..... Actually, if we compare the passage of temperature shown in Figure 38 for the N -th year with the passage of temperature for the $N + \frac{1}{2}$ year, we will see that, despite the earlier passage of air temperature through zero degrees in the N -th year in comparison with the $N + 1$ year, the amount of degree-days of frost from the time of passage of air temperature through zero degrees to the appearance

of ice in the $N + 1$ year, will be so great that it stipulates the earliest appearance of ice in the $N+1$ year. The amount of degree-days of frost in the N -th year is 16.8; the period of the appearance of ice in the N -th year is 4 January; the amount of degree days of frost in the $N + 1$ year is 1 January.

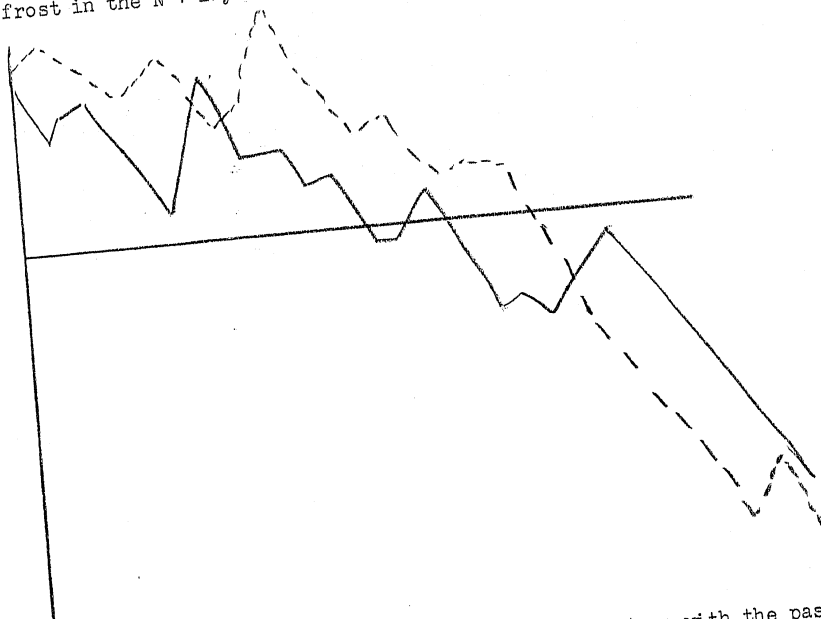


Figure 38. The changes in mean daily air temperature with the passage of time in N and $N + 1$ years.

The periods of the passage of air temperature through zero degrees and the periods of the appearance of ice are indicated with arrows.

If we also bear in mind that, as the observations have shown, the amount of degree-days of frost can be considered as a value proportional to the heat loss by a sea, then it will become apparent that it is necessary to consider the passage of the temperature

before the appearance of ice.

If the problem of the observation is not to get a closed thermal equilibrium, then in computing its elements, we can sometimes use the simpler formulas as compared with those examined above, but fully conforming with the physical substance of the examined processes. Ya. A. Tyutnev with the aid of V. V. Shuleykin's empirical formulas computed the amount of heat lost by a sea in evaporation, the heat exchange and radiation for the time between the passage of air temperature through zero degrees and the appearance of ice. The computation is made with the following relationships:

$$q_{\text{evap.}} = 48.2 V \Delta$$

$$q_{\text{T.O.}} = 30(t_w - t_a),$$

where $q_{\text{evap.}}$ is the amount of heat lost in evaporation of 1 square centimeter of sea surface in 24 hours expressed in small calories; $q_{\text{T.O.}}$ is the amount of heat in calories per square centimeter lost in a heat exchange and radiation; V is the wind speed in meters per second, averaged for 24 hours; Δ is the humidity deficit in millimeters, averaged for 24 hours; t_w is the water temperature in degrees, averaged for 24 hours; t_a is the air temperature, averaged for 24 hours.

The results of the computations made with these formulas are shown in Table 10.

TABLE 10

Year	Number of days from the passage of t_a through 0 degrees to the appearance of ice = n		$\sum_{o = C_{T.O.}}^n C_{T.O.} = \sum_{o = C_{Evap}}^n q_n = \frac{C_{evap}}{Q_{T.O.}} = \frac{Q_2}{q_n + Q_{T.O.}} \cdot Q_1 = \frac{Q_1 - Q_2}{Q_1} \cdot 100$
Astrakhan'			
1932	7	1351	1373 1.01 2724 3240 6
1933	1	567	439 0.77 1003 1800 44
1935	10	1582	1397 0.88 1979 3600 18
1936	3	469	290 0.60 759 900 15
1937	3	360	457 1.27 817 906 10
1938	4	770	519 0.67 1289 1620 20
1939	1	95	104 1.09 199 270 27
1945	3	980	803 0.81 1783 1250 21
1946	3	676	610 0.91 1286 1350 5
Gur'ev			
1932	2	449	418 0.93 867 1020 15
1934	1	375	490 1.30 865 780 11
1935	2	538	356 1.00 1073 1260 15
1946	2	212	280 1.32 492 600 18

In Table 10 in column Q_1 we have the amount of heat in gram-calories lost by 1 square centimeter of a water surface in cooling down from the temperature it had on the day the air temperature passed through zero degrees to the freezing temperature. This amount was computed by the relationship

$$Q_1 = cH(t_{w_0} - t_{w_3})$$

in which C is the heat capacity; t_{w_0} is the water temperature at the time of passage of air temperature through zero; t_{w_3} is the freezing temperature of the water and H is the mean sea depth in the area for which the computation is being made.

From Table 10 it follows that Q_1 and Q_2 are close to one another although in all cases $Q_1 > Q_2$. However, the relationship $\frac{Q_1 - Q_2}{Q_1} \times 100$ indicates that the difference varies within comparatively narrow limits and on an average is about 15 percent. The large percentage (44 percent) refers to 1933, and is explained by the displacement of the thermal field usual in such cases, caused by abundant snowfalls occurring during the period for which the elements of thermal equilibrium were computed.

This gives us the right to assume approximately that Q_1 is proportional to Q_2 or $Q_1 = k_1 Q_2$. In addition, it is apparent from the table that the relationship $\frac{Q_{\text{evap.}}}{Q_{\text{T.O.}}}$ is close to one and changes within comparatively narrow limits from year to year. The latter condition enables us to consider:

$$Q_2 = Q_{\text{evap.}} = Q_{\text{T.O.}} \cong k_2 Q_{\text{T.O.}}$$

In turn, in the relationship $Q_{T.O.} = 30(t_w - t_a)$ the quantity t_w changes within comparatively narrow limits, in contrast to t_a . This allows us to assume approximately that t_w is constant and the amount of heat lost by the water in a heat exchange from the time of the air temperature passing through zero to the appearance of ice is proportional to the sum of negative air temperatures:

$$Q_{T.O.} \cong k_3 \sum_0^n (-t_a)$$

Consequently, there will be no large error, if we write

$$Q_1 \cong k_1, Q_2 \cong k_2 Q_{T.O.} \cong \sum (-t_a)$$

Q_1 , the heat reserve of the sea in each concrete off-shore area from year to year, changes within wide limits and depends mainly on how high the water temperature was at the surface of the sea at the time the air temperature passed through zero degrees, that is, in the relationship $Q = CH(t_{w_0} - t_{w_3})$ only the quantity t_w , changes considerably, but the quantity t_{w_3} changes little depending on the salinity.

Consequently, if the above mentioned considerations do not lead to an inordinate agglomeration of errors for each concrete area, then the relationship

$$\sum_0^n (-t_a) = f(t_{w_0})$$

should be true, that is, the greater the sum of the negative air temperatures after passing through zero degrees, the greater was the water temperature at the time the air temperature passed through

zero degrees. In the more general relationships applicable in most off-shore areas, the mean sea depth for this area will figure in the form of a changing quantity, that is

$$\sum (-t_a) = f_1(Ht_{w_0})$$

Tyutnev built a relationship between the sum of degree-days of frost and the water temperature for four points located on the Caspian Sea. This relationship is graphically represented on Figure 39. As we can see, the lines of regression shown in this figure have various angles to the abscissa. A comparison of the mean depths for each of the studied areas showed that the tangent of the slope angle was proportional to the depth in this area. The less the mean depth, the less the slope angle of the lines of regression. In other words, as it should be for the relationship described above, the less the sea depth, the less the sum of degree-days of frost we need for an appearance of ice. The corresponding recomputation allowed us to form a linear equation general for all four points of the Northern Caspian, fully suitable to the equation written above. Constructing such an equation proves that the assumptions made above do not create to large unpermissible errors (Figure 40). [See Figure 39, 40 on following page]

In this manner the products of observations gave us the opportunity to study the sums of negative temperatures as an indicator of the general heat losses by a water surface. Therefore, in all those cases when Q_1 may be computed by the formula $Q_1 = CH(t_{w_0} - t_{w_3})$, when t_{w_3} changes at once throughout the entire depth or, in other words, the vertical temperature stratification is missing, for fore-

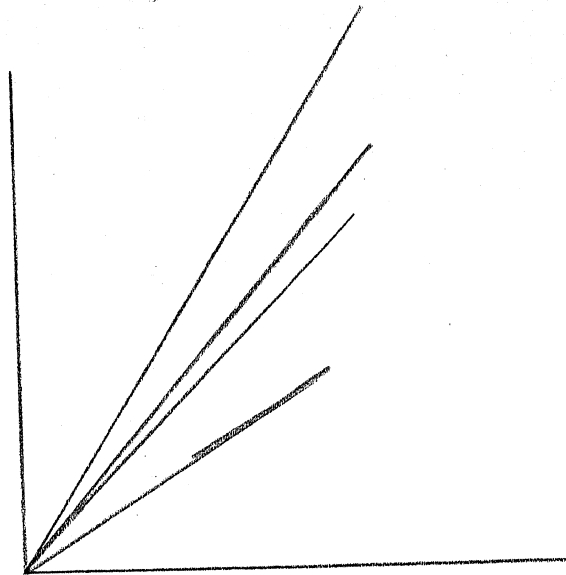


Figure 39. Relationship between the sum of negative air temperatures necessary for the appearance of ice and the water temperature observed on the day the air temperature passes through zero degrees.

1- Astrakhan', 2 - Gur'ev, 3 - Zelenga, 4- the greenhouse business.

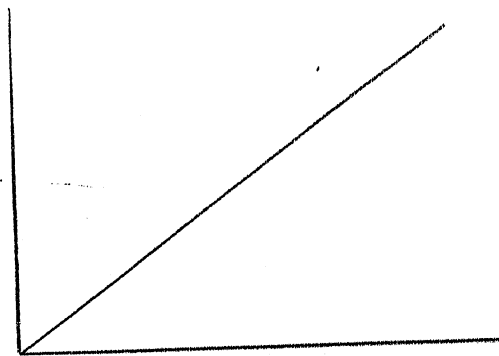


Figure 40. Relationship between the sum of negative air temperatures necessary for the appearance of ice and the heat content of water on the day the air temperature passes through zero degrees.

casting we can use the simple equation:

$$\sum (-t_a) = f(Ht_{w_0})$$

Making forecasts of periods of the appearance of ice with the aid of this relationship and forecasts of mean daily air temperatures for several days in advance brings us to the following.

From the graph on Figure 40 we determine which sum of degree-days of frost corresponds to the water temperature for which the passage of air temperature through zero degrees occurred, and from the forecast of mean daily air temperatures made for several days ahead, we compute to which date the corresponding sum of negative air temperatures will build up. The date to which the necessary sum of negative temperatures will build up is taken from the graph and taken as the date of the appearance of ice.

It is quite apparant that a simple distribution of the temperature (the vertical and horizontal temperature stratification is missing) is an exceptional case. As a rule this simple relationship must be fairly completed due to the necessity of computing the complete factors affecting the heat exchange process between the sea and atmosphere.

An example of an area where a horizontal and vertical water stratification is clearly defined may be Rzhskiy Bay which has a considerable water stratification in the vertical as well as horizontal direction.

Clearly, in a case of stratified water the surface water temperature taken by observations at an off-shore area cannot be characteristic of heat reserve of waters up to the time the air tempera-

ture passes through zero degrees in the area for which a forecast is being developed for the following reasons:

(1) Heat reserve as well as stratification and consequently also the conditions of thermal convection caused by the cooling of the surface layers of water change from year to year within wide limits.

(2) In the case of horizontal and vertical stratification of waters their heat reserve in the area for which a relationship is being developed from the time the air temperature passes through zero degrees to the appearance of ice can change within wide limits as a result of vertical and horizontal mixing of waters.

Unfortunately, it is very difficult to select the data whose use could form the stratification characteristic of waters in any area at the time the air temperature passes through zero degrees for many years and to evaluate the influence of peculiarities of water temperature distribution and salinity each year separately on the formation of ice. The absence of this data forces us to try to evaluate the influence of vertical as well as horizontal mixing by other means.

As examples of other means of computing we have the equations based on the close relationship existing ~~between current and variations~~ in sea level. As was mentioned above, turbulence increases with the growth in current speed, at the same time large current speeds are accompanied by large deviations of sea level from a condition of equilibrium. That is why turbulent mixing may be considered as an index of the absolute value of the deviation figure of a

level from the mean value from the time the air temperature passes through zero degrees to the appearance of ice. A comparison of this quantity with the sum of degree-days of frost necessary for the appearance of ice showed that for large variations in level a large sum of degree-days of frost necessary for the appearance of ice is required. The coefficients of correlation between the compared quantities reached 0.81 for the area Mersrags, and for the area Aynazhi--0.66. For a third point being studied the relationship was weak since this area, located in Irbenskiy Strait, thanks to its geographical position, was subject to a great influence from heat advection attributed to the movement of the waters of the Baltic Sea into the Gulf of Riga through Irbe Strait. The influence of heat advection coming from the Baltic Sea, in this case, evidently, is considerably larger than the influence of turbulent mixing.

However, the relationship between the sum of degree-days of frost and the sum of absolute values of deviation of the level from the mean, computed from the time the air temperature passes through zero degrees to the time of the appearance of ice, had no forecasting significance. Actually, if we use this relationship in forecasting then we would have to clarify, before computing the deviation of sea level from the mean, just exactly when the ice will first appear and what the variations in level will be from the passage through zero degrees to the appearance of ice. Of course, the problem is insolluble in this situation.

It is natural to assume that the waters 5-6 days prior to the appearance of ice are sufficiently well mixed vertically in view of intensive convection caused by the cooling of the upper layers of water. Therefore, we can disregard the computations of

turbulent mixing for an interval of time equal to 5-6 days prior to the appearance of ice, and consider the variations in level only from the time the air temperature passes through zero to some time 5-6 days before the appearance of ice.

It is natural to consider this hypothesis justified if the relationships built on this assumption are sufficiently close. The study of this problem confirmed the possibility of limiting the computation of the level slope from a position of equilibrium from the time the air temperature passes through zero degrees up to 5-6 days before the time the ice appears.

In addition to the studied peculiarities in the pattern of the Gulf of Riga the necessity of computing the thaw which quite frequently reoccurs in this area is extremely characteristic. First of all to compute this peculiarity it is necessary to make an entirely objective determination of the steady passage of air temperature through zero degrees. For such areas as the Gulf of Riga in determining the passage of t_a through zero degrees we must make a parallel computation of the sum of degree-days of frost and the degree-days of heat and use only those passages through zero degrees when the sum of degree-days of frost observed after its passage through zero degrees was larger than the sum of degree-days of heat observed from the day from which the computation of the sum of degree-days of frost was begun.

It appeared that the thaws in the Gulf of Riga, computed in a similar manner, substantially increase the sums of degree-days of frost necessary in the appearance of ice. Therefore, in the equations used in the computation of the sum of degree-days of

frost we included the sum of degree-days of heat. This equation has the following form:

$$\sum_0^n (-t_a) = f \left[t_{w_0}, \sum_0^n (+t_a), \sum_0^{n-5} \Delta h \right].$$

Here t_a is the air temperature; t_{w_0} is the water temperature; Δh is the slope of sea level from a condition of equilibrium.

The certainty of non-appearance of errors in forecasts for ± 2 days, by the equation of the above type, for three points in the Gulf of Riga of which, when analyzed, was 100 percent.

The forecast in this case as in the case of simpler equations applied for the northern part of the Caspian Sea is made as follows: The surface water temperature (t_{w_0}) is determined corresponding to the air temperature in passing through zero degrees. The sum of positive temperatures $\left(\sum_0^n (+t_a) \right)$ after the air temperature passes through zero degrees is computed, and the value of the level slope from a condition of equilibrium $\left(\sum_0^{n-5} \Delta h \right)$ is also computed. For this purpose $\sum_0^{n-5} \Delta h$ is in the beginning taken as equal to zero. The computation of $\sum_0^n (-t_a)$ is made by the equation

$$\sum_0^n (-t_a) = f \left[t_{w_0}, \sum_0^n (+t_a) \right]$$

If in this computation $\sum_0^n (-t_a)$ is so large, that in accordance with the forecast of mean daily air temperatures it will appear that more than 5 days must pass until the appearance of ice, then in this

case we must begin the computations with the quantity $\sum_0^{n-1} \Delta h$, and carry them out for the entire scope. For the date of the appearance of ice we use the day when the air temperature computed by the observations and by the forecast is close to the sum of degree-days of frost, computed by the formula developed for each separate point.

29. COMPUTATIONS FOR FORECASTING THE APPEARANCE OF ICE ON AN OPEN SEA

In the previous section we noted that the heat reserve of waters in off-shore shallow areas can be computed by the relationship:

$$Q_1 = c H (t_{w_0} - t_{w_3}).$$

Here Q_1 is the amount of heat which must be given off by one square centimeter of water surface; c is the heat capacity of water; H is the sea depth; t_{w_0} is the temperature of the water when the air temperature passes through zero degrees; and t_{w_3} is the temperature at which the water freezes.

Further, in constructing the equation for each separate off-shore area the quantity t_{w_3} was assumed constant and as a constant was computed by empirical coefficients. Unfortunately, this simplification is applicable only when the relationship $\sum_0^n (-t) = F(t_{w_3})$ is constructed for each separate area of a shallow zone. Constructing a relationship of the type $\sum_0^n (-t_a) = F(H, t_{w_0}) = F_1(Q_1)$, general for several sea areas, is possible only in cases of limited depths and little changing salinity. As we know, the freezing temperature of sea water changes with a change in salinity. To compute the freezing temperature we may use oceanographic

tables or the relationship:

$$\hat{t} = -0.003 - 0.0527S - 0.00004S^2 - 0.0000004S^3,$$

where $\hat{t}$ is the freezing temperature in degrees centigrade, and S is the salinity expressed in promilles. From the data shown below it is apparent how significantly the freezing temperature of sea water changes with a change in salinity.

S°/oo - 0	5	10	15	20	25	30	35	39
$\hat{t}^{\circ}$	0.00	-0.27	-0.53	-0.80	-1.07	-1.35	-1.63	-1.91 - 2.4

An examination of this data shows that it is not always convenient when constructing relationships between a sum of degree-days of frost and the amount of heat which is lost by water up to the beginning of the formation of ice to make a count of the degree-days of frost from the time the air temperature passes through zero. In this case the sum of degree-days of frost can build up greatly and sufficiently for freezing, but freezing will not take place in view of the necessity of having a rather low temperature for freezing sea water. Therefore, it is best to make the computations of the sum of degree-days of frost for waters of high salinity from the time the air temperature passes through values close to the freezing temperature.

The influence of sea depth on the errors which may appear as a result of disregarding the changeability of freezing-temperature in relation to the salinity of water is also quite obvious.

Actually, if we compute the heat reserve of water in an area with 20 meters depth, assuming that freezing will take place at a water temperature of about zero degrees, and begin the computation

from the time when the water has a temperature of 5 degrees, then the heat reserve will be 10,000 calories. If under the same conditions we assume that the freezing temperature of sea water is -1 degree, then the heat reserve will be 12,000 calories. The error caused by a disregard of the freezing temperature of sea water is equal to 2,000 calories.

Now let us assume that the sea depth is twice as great -- 40 meters, then, as it will be easy to compute, the error also doubles, that is, it will be equal to 4,000 calories.

This example indicates the inadmissibility of disregarding the freezing temperature of water in cases when the problem of the formation of ice over large sea depths is being explained. Computing the heat reserve in the case of a deep sea is considerably more complicated than for shallow areas. For shallow places, in many cases, we can make errorless assumptions during a period preceding the formation of ice that the temperature as well as the salinity are homogenous throughout the entire depth. In deep areas, however, the temperature and the salinity change with depth may be unique for each area.

Therefore, we cannot use the given formula for deep-water areas without preliminary computations if only because it is uncertain what depth the water mixing resulting from the cooling of the surface layers of water will reach and it is also uncertain what the water temperature is at which freezing will take place. The fact is that as far as the cooling of surface waters is concerned they become heavier and fall to the lower layers where the temperature as well as the salinity is greater. A mixing of the surface

waters with the subsurface ones takes place. The deeper the mixing creeps through the greater is the salinity of the surface waters, and consequently, the lower becomes the freezing temperature. The mixing stops only when it reaches some layer having such a high salinity that its density is greater than the density of the mixing layers of water at a temperature that has sunk to the freezing temperature.

As a result of this situation it is easy to make calculations of all the values which we must know for computing the heat reserve. To avoid large errors in these computations it is necessary to use data of deep water hydrological observations with changes in temperature and salinity at levels removed from one another by not more than 5 meters. The hydrological observations must be made not more than 1 to 1.5 months before the beginning of the ice formation. It stands to reason that the error will be as little as the period between the making of hydrological observations and the beginning of the ice phenomena. It is easiest to follow through these computations in an arbitrarily taken example (Table 11). [See table 11 on following page].

Explanation of Table 11

In column 1 we have the sea depth in meters; 2 - water temperature in degrees; 3 - salinity in promilles; 4 - conditional specific volume; 5 - freezing temperature of water computed from the salinity in column 3; 6 - mean temperature taken at two adjacent levels; 7 - mean salinity of two adjacent levels; 8 and 9 - corresponding temperature and salinity resulting from the successive mixing of the layers of water from the surface to the deeper levels. The quantities in column 8 and 9 were computed by the formula:

TABLE 11

H	t°	S°/oo	V _c	∠	T _{sr}	S _{sr}	t _{per}	S _{per}	∠ ₁	V _{c1}	t°	t-t ₃
1	2	3	4	5	6	7	8	9	10	11	12	13
0	7.10	34.00	74.05	-1.85							-1.86	8.96
5	7.10	34.00	74.05	-1.85	7.10	34.00	+7.10	34.00	-1.85	73.34	-1.86	8.96
10	6.50	34.15	73.85	-1.86	6.80	34.07	6.95	34.03	-1.85	73.27	-1.86	8.36
15	6.25	34.20	73.80	-1.86	6.37	34.17	6.75	34.08	-1.85	73.27	-1.86	8.11
20	6.10	34.25	73.74	-1.87	6.17	34.22	6.61	34.11	-1.86	73.27	-1.86	7.96
25	5.13	34.50	73.44	-1.88	5.61	34.37	6.41	34.16	-1.86	73.19	-1.86	6.99
30	1.12	35.00	72.71	-1.91	3.12	34.75	5.86	34.20	-1.86	73.19	1.12	0.00
35	0.01	35.20	72.49	-1.92	0.56	35.10	5.32	34.30	-1.87	73.11	0.01	0.00
40	-0.53	35.25	72.43	-1.93	-0.26	35.22	4.43	34.40	-1.87	73.04	-0.53	0.00
45	-1.55	35.50	72.20	-1.94	-1.04	35.37	3.82	34.58	-1.88	72.95	-1.55	0.00
50	-1.70	36.00	71.81	-1.97	-1.62	35.75	3.27	34.70	-1.89	72.80	-1.70	0.00

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$$t_{nep} = \frac{\sum t_{sr} \Delta H}{\sum \Delta H}; \quad S_{nep} = \frac{\sum S_{sr} \Delta H}{\sum \Delta H}$$

in which t_{sr} and S_{sr} are taken from the sixth and seventh columns, and ΔH is the distance between the levels in meters.

In the given case $\Delta H = 5$ meters.

In column 10 we have the freezing temperature of water computed from the salinity in column 9, that is, from the salinity obtained after the mixing of water. In column 11 we have the conditional specific volume of water computed from the same salinity and freezing temperature. In column 12 we have the temperature distribution which can be expected after the mixing of water caused by winter cooling.

The results of the computations, shown in Table 11, make it possible to obtain all the necessary elements for computing the depths which the mixing will reach and the amount of heat which must be released by the sea surface before the formation of ice as a result of the winter cooling.

Actually, let us suppose that as a result of the heat loss of a water surface the water temperature in a layer 0 - 5 meters decreased to the freezing temperature - 1.85 degrees. As a result of this specific volume of the water layer decreased from 74.05 to 73.34 (see columns 5, 4, 11), that is, the water of this

level was heavier than the water at levels of 10, 15, 20, 25 meters (only at the 30 meter depth is the specific volume 72.71 less than the specific volume 73.34, which is explained from a comparison of the data in column 4 and 11). Consequently, all the water lying above the 30 meter level is mixed. The salinity of the mixing water in the entire layer from 0 to 25 meters will be equal to 34.16 (see column 9). The freezing temperature at this salinity will decrease to -1.86 degrees (see column 10), and the specific volume corresponding to a salinity of 34.16 and a temperature of -1.86 degrees, will be equal to 73.19. This specific volume is greater than the specific volume of the water at the 30 meter level (see column 4), consequently, the convection mixing will not include the 30 meter level. A further convection mixing will be possible only when the salinity of the water in the layer 0 - 25 meters increases for any reason. An increase in the salinity may take place as a result of the formation of ice. Inasmuch as the salinity of ice is less than the salinity of the water from which it was formed, during the formation of ice a salting of the remaining water takes place. As a result of this we frequently get a continuation of convection.

In this manner, to determine the depth of water mixing caused by the cooling of the water to freezing temperature we need only to construct the graphs of the change of the conditional specific volume with depth from the data indicated in columns 4 and 11. The points of intersection of these curves will indicate the depth to which the mixing will reach.

In column 12 we have the water temperatures which should be observed on the examined vertical when the mixing reaches the limiting

depth. Obviously from the surface to this depth the temperature will equal the freezing temperature computed from the salinity obtained after the mixing to the level located between 25 and 30 meters. At the below indicated levels the water temperature remains unchanged. Therefore, in column 12 the temperatures beginning from the 30 meter level are taken from column 2.

The amount of heat which will be released through the water surface before the full convection mixing is easy to get by computing the differences between the temperatures fixed in column 2 and 12, that is, the temperatures which are observed prior to cooling and mixing and after mixing and cooling. Inasmuch as the mixing in this case reaches only about the level of 25 meters, the difference of the mean water temperature of the entire layer from the surface to 25 meters will be 8.22 degrees. To obtain the heat, expressed in calories, which will be released to the sea surface it is best to represent the quantities in the relationship:

$$Q_1 = c H (t_w - t_z)$$

here H will be the maximum depth which the convection will reach.

Examining Table 11 one may get the impression that the freezing temperature changes little. Actually, the freezing is equal to -1.85 degrees and differs altogether only 0.01 from the freezing temperature computed from the salinity obtained as a result of mixing (-1.86 degrees). However, such a small change in freezing temperature can occur only at a place of comparatively little difference between the salinity of the surface layer and the salinity of deeper layers. In the opposite case, however, the freezing temperature after mixing may be substantially different from the

freezing temperature of the surface water. Especially sharp changes in freezing temperature may be observed in those sea areas into which the waters of big rivers empty. In these areas considerable changes in freezing temperature usually take place as a result of mixing and as a result of salinity changes along the sea surface according to the distance out from estuary regions. In a manner similar to that described above, V. S. Nazarov developed over 300 hydrological stations in the seas of the Soviet Arctic. As a result of this work he computed the temperature at which the formation of ice begins. The chart made by him shows that in off-shore areas where the distillation of the sea water by rivers is especially great the freezing temperature varies from -0.8 to -1.0 degrees, at the same time that it reaches -1.8 , -1.9 degrees in sea areas.

Similarly, with the aid of the computations of the above described method we should construct charts of the heat reserve of the water which the sea must lose before the appearance of ice. In constructing the charts of heat reserve it is extremely important that the hydrological observations used for this purpose were made simultaneously. This requirement, often not so essential for constructing temperature charts at the beginning of the formation of ice, is definitely obligatory in constructing charts of heat reserve. The trouble is that in the general case the closer in time the observation is made to the period when the ice appears, the lower will be the water temperature and consequently the heat reserve.

Simultaneous hydrological surveys can be made only in very suitable instances, since many ships must be utilized for this purpose. Therefore, in deciding on the heat reserve of a sea we must

utilize the observations made on the standard (regularly repeating) hydrological cross-sections.

Using the observation data on these cross-sections for a long range period one computes the mean long range value of heat reserve for each period of time for which this computation can be made. So, for example, if the cross-sections are made monthly then the heat reserve is computed for each month of the year, but if the cross-section is made each 10 day period then this computation is made for each 10 day period .

When the cross-sections from year to year are made not exactly in the same periods, we must reduce the hydrological observations to the same date. For this date we use that on which most of the other observations were made. It is necessary to try to make the hydrological observations on a cross-section from year to year exactly on the same dates. This is not only a requirement for the convenience of working out the products of observations but also one for accuracy as each reduction of the observations to the given periods introduces a distortion in the actual chart of the phenomenon. After reducing the observations to a definite date one makes the computations of freezing temperature and heat reserve expended before the appearance of ice.

The computations are made for each vertical, completed each year separately. After that the freezing temperature and the heat reserve are averaged for each hydrological station separately for a period of many years. In this manner the norms of heat reserve are obtained. Further, if the cross-sections were made not less than once a month and the averaging is done for a sufficient number

of years, by interpolating we can compute the conditional norms for each day.

Using the norms of heat reserve and freezing temperature for a series of standard cross-sections evenly illuminating the sea we can construct the approximate charts of freezing temperature and heat reserve norms, using the hydrological stations at which the observations were made episodically. Such charts should be constructed for a separate year too when the observations are available. Just as the mean long range charts of heat reserve, the charts representing the heat reserve of the sea for separate years must be carefully analyzed and compared with the observations of ice conditions at sea and at off-shore stations. In this it is necessary to bear in mind the possibility that there may be no relation between the charts of freezing temperature of water and heat reserve and the observations of ice. This lack of relationship may be due to several reasons of which the following should be kept in mind first of all.

It may be that in spite of a large heat reserve we observed a very early appearance of ice. In this case it is possible that the ice was brought from other sea areas or carried out by a river. These phenomena are observed rather frequently. Especially frequent are the instances when the ice is carried out by the rivers.

Conversely, we can often observe the phenomenon that for comparatively small heat reserves and high freezing temperatures, computed by the above method, in one or the other area the ice will appear at a very late date, and sometimes not at all. This phenom-

enon can be observed in connection with the ensuing influx into the examined area of warm waters which, of course, can not be computed by the above described method. Besides this, this phenomenon can also be observed because this computation does not consider the mixing of waters which may take place from storms frequently acting with special force just in the period preceding the formation of ice. The mixing of waters caused by storm winds may also be approximately computed with the aid of the following concepts.

Mixing takes place as a result of the development of a large disturbance. Water particles come to a vascillating motion, after which these motions die down together while being removed from the surface of the sea to its depths and become hardly noticeable at a depth equal to half the wave length.

In the cross-section where we examined the problems connected with forecasting disturbances we indicated the methods with the aid of which we can compute the size of waves. Consequently, using the necessary relationships for forecasting disturbances, it is easy to compute the maximum wave length which was connected with this or that storm, and consequently, also, to approximate to what depth the mixing, connected with the storm activity, could reach. If it appears that the depth of the storm mixing is greater than the depth of the convection mixing then it is necessary to recompute again the heat reserve with the aid of the above described table, assuming that all the water from the surface to the depth of the storm mixing will have the same salinity and a corresponding temperature

Sometimes it may appear that the depth of the storm mixing

many times exceeds the depth of the convection mixing. Under these conditions we should always carry the calculation to the depth of the storm mixing. In many cases it may be that the depth of convection mixing is so great that there will be no need to consider the storm mixing.

Finally, in deep off-shore areas or in areas where the depth changes sharply we may observe a case where the deep layers of water come out on the surface of the sea, connected with the "running-out" wind actions. Of course in these cases the heat reserves of the sea calculated in the computations of convection mixing may be many times less. To compute the changes in heat reserves in this case, it is necessary to use the relationships of water temperature to the wind functioning in the "running-out" of the water and the emergence of the deep layers onto the surface, which can be general relationships, serving in short-term water temperature forecasts. In constructing this type of equation we can determine to what lower limit of value the surface water temperature decreased during a summer period as a result of the action of a "running-out" wind of maximum strength and duration. Knowing this temperature from the observations usually made at roadstead departure points it is not difficult to be convinced that this temperature or one close to it was observed for the entire depth of the waters at off-shore areas.

Using also the deep-water hydrological observations in this area, made at a period of time close to when the lowest possible water temperature was observed in a "running-out" wind, we can determine at which depth at sea in this period a layer of water sets in with a temperature equal to the lowest possible water temperature

on the surface in a strong and continuous "running-out" wind. It becomes apparent that the decrease in water temperature during the time of the "running-out" was related to the emergence on the surface of just this water.

Having determined in this manner the depth from which the water emerges onto the surface, we can establish whether this water lies below the level of the convection mixing or above it. If the depth from which the water emerges in a "running-out" wind action becomes less than the depth of the convection mixing, then, to determine the heat reserve of the sea, we need no supplementary computations. In the opposite case it is necessary to compute the heat reserve considering the depth from which the water emerges in sgonno-nagonnye winds. In many cases the influence of the heat reserve related to the sgonno-nagonnye wind action may be so great that the ice may not form at all in "running-out" winds. In such areas, in winter time, land floes will generally not form, and the floating ice brought by the wind or formed on the spot in wind of a "running-up" direction or in a calm are frequently carried away in "running-out" winds. In these areas we frequently have tidal leads which sometimes attain great dimensions.

All the above material shows that before starting to develop ice forecasts for deep-water areas it is necessary to study carefully the hydrological pattern of these areas and to work hard at providing forecasters with the products of observations.

30. FORECASTING THE APPEARANCE OF ICE IN THE OPEN SEA

Using the products of observations and the preliminary computations (see Section 29), we can begin working on ice forecasts

for deep-water sea areas. For several sea areas in which the air temperature at off-shore areas is considered representative we may use the relationship between the amount of heat expended by the sea before the appearance of ice and the sum of negative air temperatures. This relationship has the form:

$$\sum(-t_a) = F[H, (t_{w_{SR}} - t_3)]$$

In the case of a deep-water sea area instead of H we must use the sea depth corresponding to the full convection mixing occurring in the cooling of a sea surface, or the depth of the greatest mixing occurring as a result of storm action, or the depth of mixing corresponding to the layer of water emerging on the sea surface in the case of "running-out" wind action. The selection of the corresponding depth is made depending on the sea area (if for a given area one or the other of the named depths is characteristic) and depending on the preceding and expected meteorological situation.

For example, if the depth of the convection mixing in the examined area is not very great, but in the preceding time in the given season of ice appearance there was no stormy weather and no such increase in wind is expected which could mix the water to a depth deeper than the depth of the convection mixing, then in forecasting we should take only the depth of the convection mixing in the computing formula. Or if, despite the observed and expected then the computing formula should take the depth of the storm mixing when it is greater than the depth of the convection mixing. Or, finally, if the depth of the stratification of waters emerging onto the surface in "running-out" winds is greater than the depth of storm and convection mixing and we expect a development of "running-out" winds before the appearance of ice, then it is necessary to take

the depth corresponding to the "running-out" phenomenon in the computing formula. But besides the depth the water temperature is also included in the computing formula (mean for the layer from 0 to H meters) as is the freezing temperature. The freezing temperature in a majority of the cases can be taken as the mean long range one corresponding to the accepted depth of mixing, and the mean temperature of the layer can be taken only from observations. Therefore, every short term forecast of ice appearance must be preceded for a short period of time up to the appearance of ice, by an observation of water temperature and salinity at the corresponding verticals. These observations are usually made either on the standard hydrological cross-sections, or from regular ships of the merchant fleet travelling regularly on very exact courses. The closer to the appearance of ice the corresponding observations were made, the more exact, of course, can the computation be made of the heat reserve of water and the more precise will the forecast of the appearance of ice be.

We noted above that this type of computation of the heat expended from the sums of the negative temperatures was possible only when the observations of the air temperature made at a shore station are sufficiently representative for a deep-water sea area. But there cannot be many such cases. Air temperatures observed only at the shores of a sea may be more or less representative for most coastal sea areas. In most of the cases, however, we cannot use the air temperature observed directly at a shore station for any characteristic of a heat condition at sea.

This occurs because, in moving over a sea in the fall-winter

period, the air heats up greatly and therefore comes to the point where a computation of the time of ice appearance is made with the temperature sharply distinguishable from that which is observed at the shore. In exactly the same way in an opposite movement of air from sea shoreward its temperature enroute also undergoes considerable changes although less so than in the first case. Therefore, in general, for an open sea it is not as simple to compute the periods of the appearance of ice as it is for a coastal area. For the former it is necessary to either set up special computing relationships, or what is considerably better, make a full computation of the thermal equilibrium of the sea surface. How a computation is made of the full thermal equilibrium was explained above in the appropriate section; here, however, we must note those peculiarities with which these computations are concerned.

When the sea is sufficiently well covered by information received from shipboard stations no special difficulties in computing the thermal equilibrium are encountered. The information received from shipboard stations is plotted on the chart together with the shore observations. Analyzing these charts and using interpolations for any point at sea, we can find the approximate values of one or the other element used in the formula for computing the thermal equilibrium.

It is considerably more difficult to obtain values more or less close to the actual values of the elements from which the computations of the thermal equilibrium are made. If there are very few observations at sea from shipboard stations or if they are completely missing we must do some very extensive preliminary work direc-

ted at establishing relationships and corrective coefficients with the aid of which we can evaluate roughly the sea condition from the data of shore observations. Inasmuch as there is no necessity of dwelling here on this rather difficult problem it remains to note that to make a proper forecast of the period of ice appearance in an open sea is just as simple as in coastal areas if we know all the necessary elements for its compilation: the depth of the mixing, the mean temperature of the water layer subject to cooling before the appearance of ice, the freezing temperature, the amount of heat expended through the sea surface from the time when the mean temperature of the layer, subject to cooling, was measured to the time the forecast is developed, and, finally, the forecast of the meteorological elements from which the thermal equilibrium of the sea surface can be computed for each day before the appearance of ice.

If all the enumerated elements are available then, for compiling the forecast of the time of ice appearance the following computations are made:

(1) The heat reserve is computed by the relationship presented above;

(2) from the heat reserve computed by the indicated method the amount of heat lost by the sea is computed from the time the hydrological observations were made to the time the forecast was compiled (this quantity of heat is computed from the actual observations at shipboard stations of the weather condition with the aid of the formula for computing the thermal equilibrium of a sea surface);

(3) from the weather forecast, with the aid of the same formulas for computing the thermal equilibrium, we compute for each

for each day the amount of heat which will be given off through the sea surface, after which the amount of heat lost by the sea for each day is subtracted from the heat reserves of the sea until the heat reserve becomes zero. The day on which we get a full loss of heat reserve by the sea is the day when we should expect the appearance of ice in the area being studied.

The formulas by which we compute the times of ice appearance in an open sea may be written in the following form:

$$Q_1 - Q_2 - q_1 - q_2 - q_3 - q_4 \dots q_n = 0$$

In this formula Q_1 is the heat reserve at the time the temperature and salinity at sea is taken; Q_2 is the heat loss by the sea surface from the time the measurements were taken to the time the forecast was compiled; q_1, q_2, q_3, q_n are the amounts of heat which will be lost by the sea surface from the time the forecast is compiled to the time the ice appears for each separate day; the indexes 1, 2, 3, etc. designate the number of (consecutive) days from the forecast to the appearance of ice. The forecasts made in the above described manner are not made at the present time because of the absence of the necessary hydrological observation.

31 FORECASTING FREEZING PERIODS

If in computing the time of ice appearance we have to consider the amount of heat which must be lost by a sea surface in order to prepare the water for freezing, that is, to lower the temperature of the water surface to the freezing temperature, then in forecasting freezing it is usually necessary to consider the latent heat which is separated in changing water into ice. Unfortunately

to compute the latent heat is no less difficult than to compute the heat reserves. The latent heat of freezing can be considered as a constant quantity.

To convert 1 cubic centimeter of water at the freezing temperature into ice it is necessary to remove about 80 calories. But what quantity of water is necessary to be converted to ice in order that the entire visible surface of a sea is covered by stationary ice under this or that meteorological conditions still remain almost unstudied.

Actually, if we imagine completely calm cold weather and, consequently, a complete absence of disturbances, then obviously, in a short while (if, of course, the temperature of the surface water reaches the freezing temperature) the entire visible sea surface will be covered by a thin crust of ice. If the water from which the ice forms has a high salinity, the crust of ice formed will be elastic; if, however, the water is fresh, this ice will be very brittle resembling glass. In either case the entire surface of the sea will be covered by an uniform solid layer of ice.

If the calm weather will hold out for a long time then the thickness of the ice may reach a considerable magnitude and then the solid stationary ice may last until the spring despite the development of storm winds later on. In the opposite case, however, an arising wind will quickly break up the thin cake of ice formed during the calm weather. Expanses of clear water will form and the formation on the clear water can also be observed as can the agglomeration of ice, a mixing of the ice as a result of the bumping together of the separate chunks of ice, and a piling up of the chunks of ice one on the other. Under these conditions to compute the amount of water which should be converted to ice to form a solid stationary coating is extremely difficult.

Besides the purely mechanical reasons preventing the formation of a solid stationary ice coating it is necessary to compute the thermal processes essentially delaying the formation of ice. As was noted above, the wind mixing in many cases can reach greater depths than convection mixing related to the cooling of water surface. Consequently, in many cases a considerable slowing up or even a full stoppage of the ice-forming processes caused by the mixing of waters may be possible.

Finally, we must not forget the salting of the waters caused by the formation of ice. As we know the salting of waters will increase its density and this may strengthen vertical convection, bring it to a greater depth, and consequently, bring to the surface new heat reserves.

All that was presented above makes it extremely difficult to compute the amount of heat which a sea must lose to form a solid ice coating on its surface. This explains why it is considerably more difficult to make forecasts of freezing on the high seas than to make forecasts of the appearance of ice. Forecasts of freezing, as a rule, are less successful in comparison with the forecasts of ice appearance.. At the present time there is not a sufficient amount of work being done to clarify this problem. That is why here we cannot give examples of freezing forecasts sufficiently general for any sea area, but must limit ourselves individual examples which allow us to give only the most general presentation of an approach to the solution of this problem by purely empirical methods.

We can consider the simplest of these examples the computation of the amount of heat which must be released by a sea surface before the foreformation of ice of a certain thickness. Using sufficiently reliable observations of the formation of ice including also the measurements of the thickness of ice on days when the entire visible water surface was covered by solid stationary ice at several points we can establish that in separate areas a solid ice will be formed with the same ice thicknesses. This phenomenon can be most often observed at the mouths of rivers and well closed harbors. If such areas are established and the ice thicknesses are found in which in an overwhelming majority of cases we observe a solid freezing, then, ~~latter~~ it is not difficult to compute the amount of heat lost by a water surface from the time of the appearance of ice to the formation of solid stationary ice.

For this purpose it is sufficient to multiply the ice thickness expressed in centimeters characteristic for the point studied by the ice density equal approximately to 0.92 and by the latent heat of melting of ice equal approximately to 80 calories. This step will determine the amount of heat which must be given off by a sea surface before the formation of solid stationary ice. Using this quantity and knowing the relationship between the sum of degree-days of frost and the thermal equilibrium of a sea surface it is easy to compute what sum of degree-days of frost must be accumulated from the time of the first appearance of ice to full freezing and consequently, to work out the forecasts of the periods of freezing from the forecasts of the mean daily air temperatures. But there are few areas for which we can establish so easily the equation of the sum of the degree-days of frost necessary for full freezing.

Usually, the sum of degree-days is variable and not constant as in the examined case.

For example, Ya. A. Tyutnev succeeded in showing what an essential part the wind pattern plays during the period between the appearance of ice and freezing. For this purpose he set up relationships of the sum of degree-days of frost accumulated in this period with the sum of wind speeds coinciding with the sgonno-nagonnyy direction (Figure 41). On Figure 41 we have shown the relationships developed for two stations located in essentially separate areas in its hydrological field. One of the stations -- Kandalaksha -- is located at the apex of the deep-water Kandalaksha Bay of the White Sea, and the second -- Mud'yug Island -- is located in the comparatively shallow Dvina Bay in the same sea. On the ordinate the sums of degree-days of frost are plotted, and on the abscissa the sum of wind speeds taken for four periods of observations for Mud'yug in sectors from west to north-northwest and from east to south-southeast and for Kandalaksha from north to west-northwest and from south to east-southeast. The sums of wind speeds were taken for the period of time from the appearance of ice to a time 5 days preceding freezing. The five days prior to freezing were omitted so that when making freezing forecasts one will not have used wind forecasts.

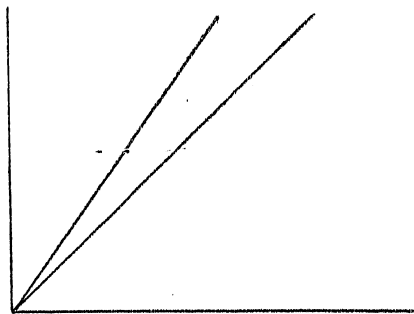


Figure 41. The relationship of the sum of negative air temperatures to the summary wind speed : 1 - Kandalaksha, 2 - Mud'yug.

As we see of the two straight lines, the first one, drawn through the points representing the relationship for Mud'yug has a smaller angle with the abscissa in comparison to the line of regression representing the relationship for Kandalaksha. This circumstance indicates that the wind for deep-water areas plays a larger part than for shallow ones.

However, the relationships presented here can be studied only as a starting phase of the development of a method for forecasting freezing. In this field the development of theoretical studies is held up by the singular complexity of the phenomenon, and the empirical studies are still very poorly supported by the products of observations in sufficiently complete quantity.

32. FORECASTING THE GROWTH OF ICE

The problem of the growth of ice is considerably better developed. Many studies have been devoted to this problem for a long time; among them it is necessary to note one of the latest theoretical studies completed by A. G. Kolesnikov. His study permits giving a correct interpretation with empirical coefficients obtained in a majority of empirical relationships between the sum of degree-days of frost and the thickness of an ice-coatings.

The growth of the ice coating occurs from the lower surface of ice. The amount of the newly formed ice is proportional to the amount of heat lost by the sea through the upper surface of the ice coating. The upper surface of ice is often covered by snow, in regards to this we must consider not only the amount of heat lost through the upper surface of ice coating but also through the

upper surface of the snow covering over the ice. It is necessary to consider the effect of the snow cover on the growth of ice, inasmuch as the snow has extremely low thermal conductivity. In this manner the snow cover protects the ice from heat loss which flows continuously from the water surface as rather warm and to the upper surface of the ice as rather cold. The ice also has low thermal conductivity which according to the studies of V. V. Shuleykin, I. I. Rusanov, and V. A. Ryabchikov is $\lambda_1 = 0.0045$. For comparison we can bring in the relationship obtained in the last century by the Russian scholar Abel'son determining the thermal conductivity of snow in relation to its density ρ_0 ; $\lambda_0 = 0.0068 \rho_0^2$. From this relationship it is easy to note that the thermal conductivity of the snow cover is considerably less than the thermal conductivity of ice. For this purpose it is sufficient to use ρ_0 as equal to 0.2, 0.3, 0.4, 0.5. The less the density of the snow cover, the less its thermal conductivity and the better the snow protects the heat, and under continuing same conditions the slower the ice grows. Fresh snow newly fallen always has a smaller thermal conductivity than lying and packed snow. In comparison with the snow cover the thermal conductivity of ice can be considered unchanging.

The loss of heat through an ice cover may be considered as ~~proportional to the difference between the upper and lower ice~~ surface, and the loss of heat through the snow cover as proportional to the difference in the temperatures on the surfaces of the snow facing the ice and the air. However, in solving the problem of the growth of ice covered by a layer of snow we can proceed even more simply assuming the heat loss is proportional to the difference in temperatures on the lower edge of the ice and the upper edge of the snow. The temperature of the lower surface of

the ice may be considered as equal to the freezing temperature of water of a given salinity, and the temperature of the upper surface of the snow can be considered approximately equal to the air temperature.

Furthermore, considering entirely on that basis/ the freezing temperature of water to be constant and bearing in mind the comparatively small difference of this temperature from zero, we may assume the heat loss by the sea through the ice and the snow cover to be dependent on the air temperature. Aiming at the greatest possible simplification of his rather complicated formula obtained by theoretical means and describing the added growth of the thickness of ice with great accuracy, A. G. Kolesnikov came to a relationship greatly resembling the empirical formulas. This simplified formula of Kolesnikov has the following form:

$$h^2 + 8.25 \delta h = 9 \Sigma - t_a$$

Here we designate the thickness of ice in centimeters by h ; δ denotes the height of the snow cover on the ice in centimeters; and t_a is the temperature of the air in degrees. N. N. Zubov's empirical formula has a similar form and is distinguished only by coefficients and by the fact that the snow cover is not considered:

$$h^2 + 50 h = 8 \Sigma - t_a$$

Even Kolesnikov considerably simplified theoretical form-

ula introduced here has a great advantage over Zubov's formula because Kolesnikov's formula ~~considers the thickness of the snow cover~~ and Zubov's does not have this important element. Besides, the constant coefficients in Kolesnikov's formula can be determined at any time depending on the local conditions with the aid of the rather complicated formula already mentioned.

For example, the coefficient shown above with h , was obtained as a result of the fact that in its computation the density of snow was taken as equal to 0.4. In the same case when there is a need, or more precisely, an opportunity to use in computations not only the height but also the density of the snow cover, this formula can be presented in the form:

$$h^2 + \frac{1.32}{\rho_0^2} \delta h = 9 \Sigma - t_a.$$

With this formula we can compute the thickness of the ice cover from the time of the formation of a thin crust on the sea surface to any point in time. For this it is necessary to know the sum of degree-days of frost for each day for which we need to compute the ice thickness and the mean thickness of the snow cover and the density for the time elapsing from the beginning of the formation of ice to the given day. The computations with this formula are inconvenient and can prove to be insufficiently accurate. It is considerably more convenient in practice to compute the growth of ice from day to day with the same formula but taken in the differential form:

$$\Delta h = \frac{a \Delta \Sigma - t}{h + \frac{b \delta}{\rho_0^2}}$$

It is convenient to use this formula for computing the growth of the thickness of ice. If the mean daily air temperatures for each day are known from the time of the formation of ice on the sea surface and if sufficiently accurate measurements of the density and height of the snow cover in the ice are available, we can carry on computations substituting in the formula the values of these quantities for each day getting for it the growth of ice as a result of the computations with the figures. In this instead of h we substitute the thickness of ice formed as a result of the growth of ice

for the previous days, that is $\sum \Delta h$.

As this formula shows, the growth of the thickness of ice for a 24 hour period increases with an increase in the negative air temperature and decreases with an increase in the sum of the thickness of ice and the thickness of snow obtained after multiplying them by the corresponding coefficients considering the value of $\frac{1}{\rho_0}$ a constant coefficient.

Inasmuch as the coefficients a and b also cannot be considered constant, it is best to select a corresponding coefficient for each area from the beginning, thereafter evaluate the errors obtained by using the known coefficients a and b, and then only use these formulas for a practical purpose, that is for computing the thickness of ice or for making forecasts.

P. P. Nikiforov in the Marine Forecasting Section of the Central Institute of Forecasting checked the applicability of this formula for the Azov Sea and obtained a rather good convergence of the computed values with the observed ones (figure 42). When checking a formula it is necessary to be extremely critical of the products of observations of the thickness of ice, since quite frequently the measurements of the thickness of ice can be nonrepresentative for the area of observation. This will be more thoroughly covered in the section on forecasting spring ice phenomena.

Using the relationship studied above the development of forecasts of the thickness of ice does not present any fundamental difficulties. It is sufficient for the forecast of the thickness of ice for this or that period of time to use the forecasts of air tem-

perature and the thickness of snow on the ice in order to compute easily the growth of ice for a future time. If, for some reason, it is difficult to obtain a forecast of the increase in the thickness of the snow cover then the computations of the growth of the thickness of the ice are made only from the forecast of the air temperature, from the thickness of ice known for a previous time and from the thickness and density of the snow cover, and in the majority of cases they give satisfactory results. Large errors probably are made only when the thickness of the snow cover increases sharply or as a result of a sharp change in the density of the snow following a thaw or other causes.

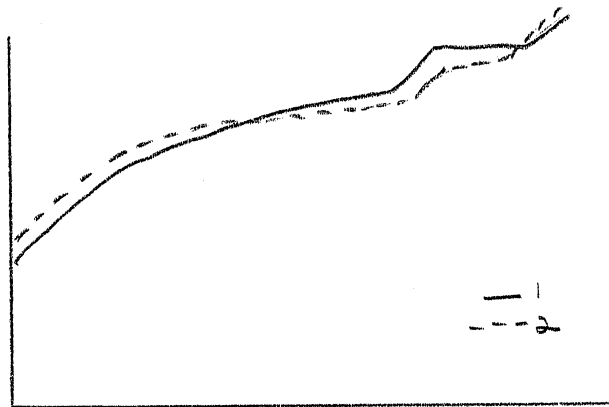


Figure 42. The thickness of ice; 1 - actual, 2 - computed.

33 FORECASTING ICE DRIFTS

The movement of ice under the action of wind, despite its tremendous importance to navigation, has not been sufficiently studied up to the present time even though several studies of a theoretical and empirical nature have been devoted to this problem. The reason for this, first of all, is that the phenomenon itself is very

complex and the making of observations of ice drift and of the factors determining and accompanying it requires a very great effort.

In the years of the conquest of the Northern Sea Route many observations were made of the ice drifts in the Arctic Ocean and the seas of the Soviet Arctic. The most complete observations, achieving world fame, are from the Severnyy Polyus and Sedov stations.

Soviet scholars widely used these observations of ice drifts and hydrometeorological conditions for the establishment of the relationships of ice drifts and wind action.

V. V. Shuleykin formed a theory of ice drifts and compared the results of the observations of the Severnyy Polyus station with the conclusions of Shuleykin's theory but at the same time considerable deviations of the theoretical computations from the observations were discovered. The basic propositions of Shuleykin's theory are presented below.

V. V. Shuleykin assumes that ice is subject to the tangential (directed along the tangent to the ice surface) wind pressure τ_a , proportional to the square of the wind speed, that is,

$$\tau_a = k_a \rho_a V^2$$

Here k_a is the coefficient of friction; ρ_a is the air density; V is the wind speed.

Under the action of this force the ice begins to move and in turn forces the water to move exactly as the wind acting on a free water surface caused drift currents. In other words, the ice moving under the action of the wind itself causes the water to move

through friction. Hence it would seem that the motion of the water under the ice should not be in any way different from the usual drift current, which, according to Ekman's theory described above, deviates 45 degrees to the right from the wind direction. However, as V. V. Shuleykin notes and as observations show, the layers of water directly adjacent to the lower surface of the ice stick to it and a turn to the right by the flow begins at some distance from the lower surface of the ice. The depth at which the flow begins to turn to the right depends on the size of the ice mass and the greater it is the greater is its size.

Besides, we must note that the entire system is in motion: the air, the ice and the water under the ice. Therefore, in computing the tangential wind pressure on the ice we should take not simply the wind speed V , but the wind speed computed in relation to the speed of the moving ice. But the wind speed in comparison with the speed of motion of the ice is so great that the relative wind speed is almost the true one.

Otherwise, the question is one of comparing the speed of motion of the ice with the flow under the ice. In this case the speeds are entirely comparable in value, and therefore, when computing the friction force τ_w between the water and the ice it is ~~necessary to take the speed of ice motion relative to the wind.~~ This speed on Figure 43 is designated by the vector r . The friction force of the ice against the water τ_w would be written in the form:

$$\tau_w = k_w \rho_w r^2$$

where k_w is the coefficient of friction between the water and the ice and ρ_w is the water density.

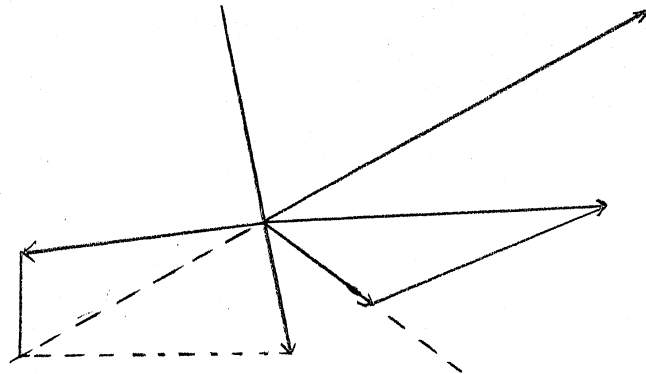


Figure 43. A diagram of the distribution of forces acting on the ice, according to Shuleykin.

As in any motion on earth in a motion of ice under the action of a wind a coriolis force develops which can be expressed as

$$K = 2m\omega u \sin \phi$$

In this equation the following designations are used: m is the mass of a column of ice with a base of 1 square centimeter; ω is the angular speed of rotation of the earth; u is the speed of the ice drift; ϕ is the latitude of the location.

In Figure 43 we have a diagram of the distribution of forces causing ice drifts and associated with them. The vector u represents the speed of the ice drift, vector $\hat{u}$ is the tangential wind pressure on the ice, ϕ is the angle at which the drift speed deviates to the right of the wind, ω is the speed of the drift flow arising under the ice, r is the difference between the vectors u and ω representing the speed of the ice drift relative to the water. The vector $\hat{U}_w$, the force of friction between the water and the ice, is parallel but opposite in direction to the relative speed r , and finally, vector K , Coriolis force developed

by the motion of the ice, is 90 degrees to the vector forming the speed of the ice drift.

The force of the tangential wind pressure on the ice, the friction between the ice and the water and the Coriolis force equalize each other in a steady movement.

$$\vec{C}_a + \vec{C}_w + \vec{K} = 0$$

Solving this vector equation, V. V. Shuleykin found similar expressions determining the angle which is formed by the direction of the ice drift with the wind direction, and also the relationship of the speed of the ice drift to the wind speed.

In Figure 44 is shown the relationship of the angle to the gradient pressure, and in Figure 45 there is represented graphically the relationship of the drift to the gradient pressure proportional to the wind speed. Analytical expressions from which curves are drawn shown in these figures are not brought in here because of their complexity.

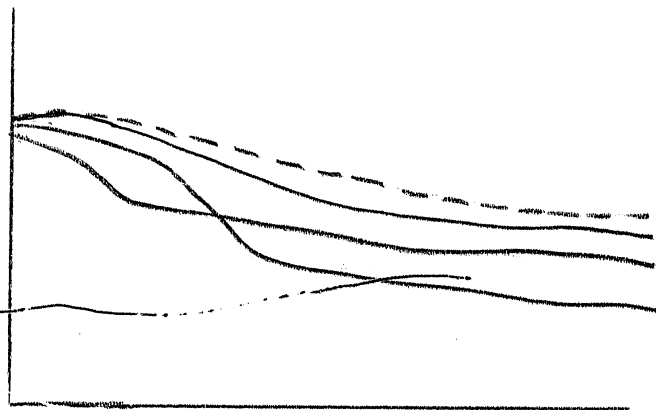


Figure 44. The change in the angle of deviation of drift from the gradient depending on the value of the gradient: a - from the studies

of Vasil'ev and Glagolev, δ - from Shuleykin, the friction coefficient is $8 \cdot 10^{-3}$, B - from Gevorkyan, τ - from Shuleykin, the friction coefficient between the ice and the water is $4 \cdot 10^{-3}$.

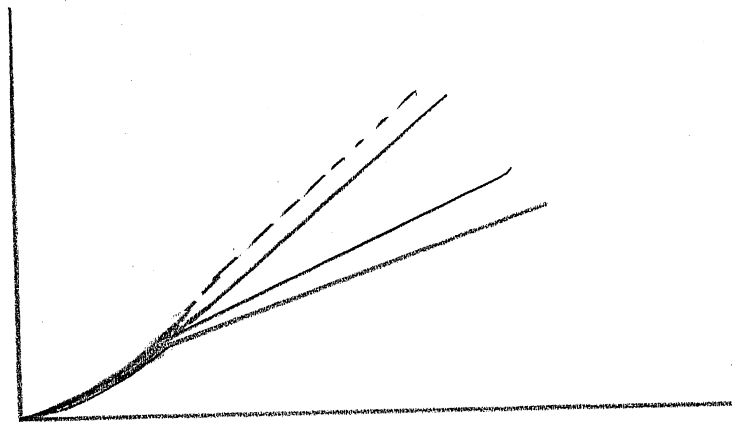


Figure 45. The change in the speed of the ice drift depending on the value of the gradient pressure: a- from the research of Vasil'ev and Glagolev, δ - from Shuleykin, the friction coefficient is $8 \cdot 10^{-3}$, B - from Gevorkyan, τ - from Shuleykin, the friction coefficient is $4 \cdot 10^{-3}$.

As it follows from Figure 44 and 45, the speed of the ice drift changes in accordance with a complex principle, but beginning with a wind speed of 3 meters per second, or 0.01 millibars per kilometer, it grows proportionally. The angle which constitutes the direction of the drift in relation to the wind direction changes within wide limits. In weak winds the deviation of the drift from the wind reaches almost 90 degrees to the right, and from the gradient pressure $90 + 80 = 170$ degrees, whereupon, because of an increase in the wind, the angle approaches its limiting value of about 33 degrees from the wind and 113 degrees from the gradient pressure.

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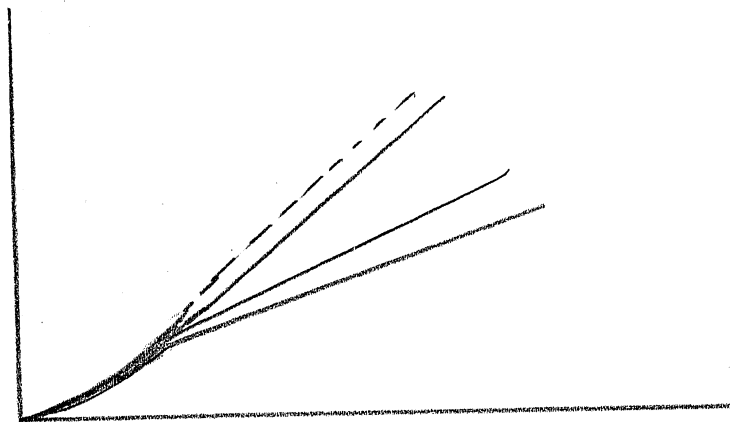


Figure 45. The change in the speed of the ice drift depending on the value of the gradient pressure: a - from the research of Vasil'ev and Glagolev, δ - from Shuleykin, the friction coefficient is $8 \cdot 10^{-3}$, B - from Gevorkyan, τ - from Shuleykin, the friction coefficient is $4 \cdot 10^{-3}$.

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Somewhat later, N. N. Zubov, analyzing the drifts at Sedov station developed the material by averaging and came to a very simple but not accurate rule, by means of which, in his opinion, one can easily compute the direction of an ice drift as well as the speed. This rule is as follows: ice drifts along an isobar of atmospheric pressure with a speed directed along the tangent to the isobar from an observer located so that the low pressure is to his left and the high pressure to his right. The absolute value of the speed is proportional to the gradient pressure, or in other words, the speed of ice changes linearly depending on the wind speed.

Being the simplest, Zubov's rule came into use by the specialists working in the field of ice forecasting in computing the ice drift; however, many were soon convinced of the inaccuracy of the computations based on the application of the rule. As a result there was a need of correcting the products of observations of Zubov's rule as well as the quantitative deductions arising from Shuleykin's theory.

In the Section of Marine Forecasting in the Central Institute of Forecasting, K. P. Vasil'ev and M. G. Glagolev worked on the establishment of a relationship between the gradient atmospheric pressure and the wind at the surface of the earth. It turned out that for areas of the Kara Sea the wind deviates 80 degrees from the gradient pressure and the wind speed is related to the gradient pressure by the following relationship: $V = 280 \Delta p$

Here V is the wind speed in meters per second, and Δp is the gradient pressure expressed in millibars per kilometer.

N. N. Zubov reckoned, relying on the development of the products of observations, that the wind deviates 28 degrees to the left from the isobar or forms an angle of 62 degrees with the gradient pressure, and the ice, drifting, deviates 28 degrees to the left from the wind direction, that is, moves exactly along the isobars. As the studies of Vasil'ev and Glagolev showed, the ice drift cannot coincide with the direction of the isobars inasmuch as the wind deviates from the gradient pressure (for the latitude of the Kara Sea) 80 degrees, and not 62 degrees as Zubov supposed, and consequently in using Zubov's rule the error is 18 degrees for this reason alone.

Further Vasil'ev and Glagolev began analyzing the products of observations of Severnyy Polyus station in an attempt to determine the degree of applicability of Shyleykin's theory to the computations of an ice drift. In formulating the theory of ice drift Shyleykin used the products of observations of the same situation; however, he made comparisons of the direction and speed of the drift with the wind speeds observed only at one point located at Severnyy Polyus station.

It is understandable that the observations of the wind speed at one point may incidentally have had the characteristic of a drift of a large mass in the Arctic Ocean, inside of which an ice field moved with a station located on it. Attempting to get a more objective characterization of the wind field under whose direction a drift flowed, carefully compiled synoptic charts were used Chkalov's and Gromov's flight over the North Pole to America. From these charts pressure gradients were taken in such a manner that the station always appeared in the center of the circle of a radius

of 350 - 400 kilometers, and the pressure gradients were computed in a direction of two mutually perpendicular diameters. The real pressure gradient was found by the vector addition principle. However, since the synoptic charts which were used were not carefully sorted out they could not be considered faultless, inasmuch as there were very few stations on the chart in this area. To define more accurately the gradients obtained in the described manner the observations of Severnyy Polyis station were taken over the wind speed and direction and from the wind the pressure gradients were computed with the aid of the relationship obtained earlier:

$$V = 280 \Delta p.$$

In this the angle between the pressure gradient and the wind was taken as 80 degrees.

The pressure gradient taken from the synoptic charts and computed from the wind speed and direction observed at the station was added and divided in half. By this averaging more probable characteristics of the wind flow acting over a large area were obtained.

Shuleykin's theory assumes that the ice drift has steady motion. In nature, however, a steady motion is very rarely observed over a long period of time. In regard to this Vasil'ev and Glagolev attempted to decrease the errors related to the unsteady pattern of the drift at Severnyy Polyus by a gradual evening out of the drift speed and the gradient pressures computed in the described manner. The evening out was accomplished by computing the sliding mean values of wind speed and pressure gradients for 4 days.

In the graphic representation of the angle of deviation of the drift from the gradient pressure it turned out that it was

30 - 40 degrees. However, there was a considerable dispersal of dots on the graph. The dots deviating greatly from the line of the relationship in an overwhelming majority of cases represented those sections of the drift where the ice field on which Severnyy Polyus station was located did not travel in a straight line but made knots. A straight-line drift was observed when the wind did not act in the general direction in which the ice field travelled for a great portion of the way.

In Figure 46, derived from the work of the cited authors, we have shown the change in the projection of the gradient pressure causing a drift in the general direction. The sections of the drift in which the ice field was observed in a straight-line motion are indicated by a conventional marking under the curve presented on the figure. In most cases these markings coincide with negative or close to zero values of the projection of the gradient pressure which in turn indicates the opposite or transverse wind direction in relation to the wind causing a general drift. By excluding the sections of drift on which the ice field made knots it was possible to find a rather clear relationship between the angle of drift deviation and the direction of the gradient pressure or, which is the same thing, the wind direction depending on the value of the gradient pressure or wind speed (Figure 47). Because of the increase in wind speed the angle between the wind and the drift decreases and approaches to some limiting value around 30 - 40 degrees. This very same regularity in this phenomenon was revealed by Shuleykin's theory.

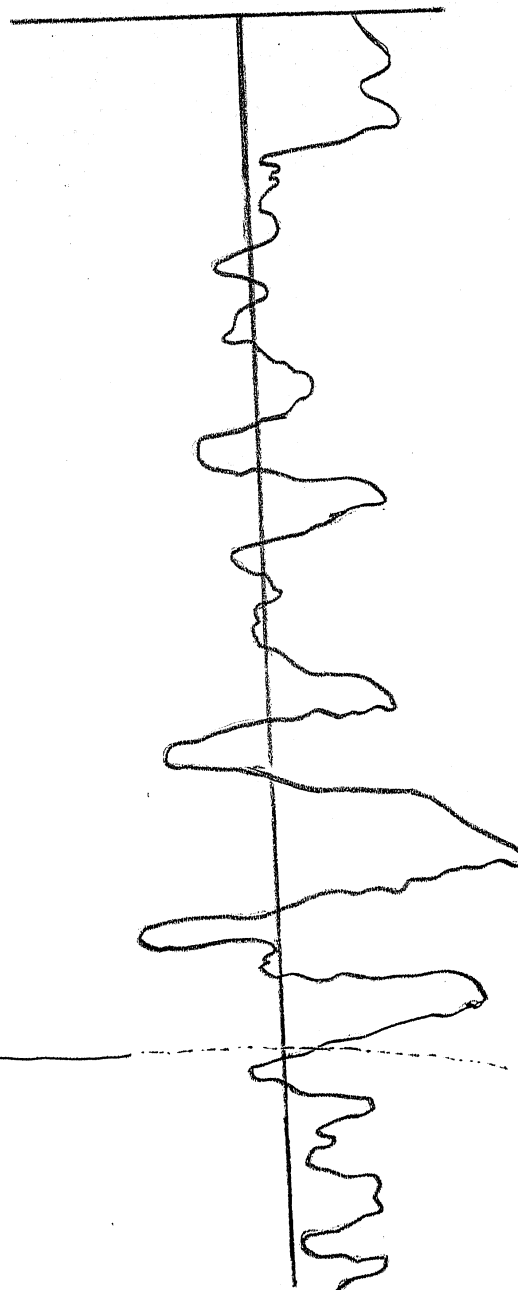


Figure 46. The change in the general direction during the projection of the gradient pressure determining a drift.

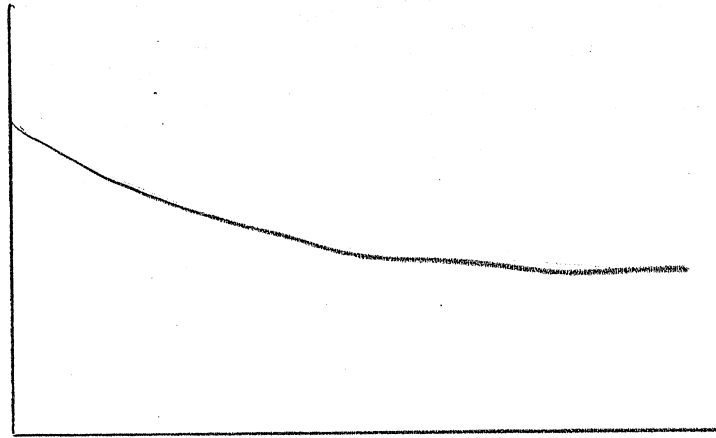


Figure 47. The change in the angle of deviation of a drift from the direction of the gradient pressure depending on the gradient value.

In a similar manner a comparison of the speed of drift ice with the atmospheric gradient pressure or, which is the same, with the wind speeds, allowed us to notice the law of the change in the speed of drift ice depending on the wind speeds or on the value of the gradient pressure (Figure 48). As can be seen easily, the curve drawn through the points does not pass through the origin of the coordinates. This indicates that even when the wind stops the movement of ice continues. The latter can be explained only by the existence of a current, not determined by the wind, acting in the area (even very large) of the drift of the station. This current can be conditionally called steady. Excluding it from the drift speed made it possible to construct relationships of the angle of deviation from the direction of the gradient pressure to the wind speed or to the value of the gradient pressure. These relationships are shown in Figures 44 and 45. Examination of these

curves and comparison with the curves derived from the work of Shuleykin (see Figures 44 and 45), show that there is no coincidence of the path of the empirical curves obtained by Vasil'ev and Glagolev with the theoretical curves even though they are sufficiently close, but, what is especially important, the path of the theoretical curves is entirely similar to the path of the empirical curves. From here it is evident that the theory of drifting ice made by Shuleykin correctly represents the main outline of this phenomenon, but at the same time there is a pressing need for further determining the coefficients which enter into the relationships of the angles of deviation of drift and the drift speeds to the wind speed or the value of the barometric gradient. Until the precise coefficients are found we should use the relationships described here. These relationships represented graphically in Figures 44 and 45 are similarly expressed as follows:

(1) The angle of deviation from the wind is determined by the formula:

$$\tan \psi = 0.30 + \frac{5.0}{u}$$

(2) The drift speed u depending on the wind is determined by the formula:

$$u = 0.22 \sqrt{\cos \psi} V$$

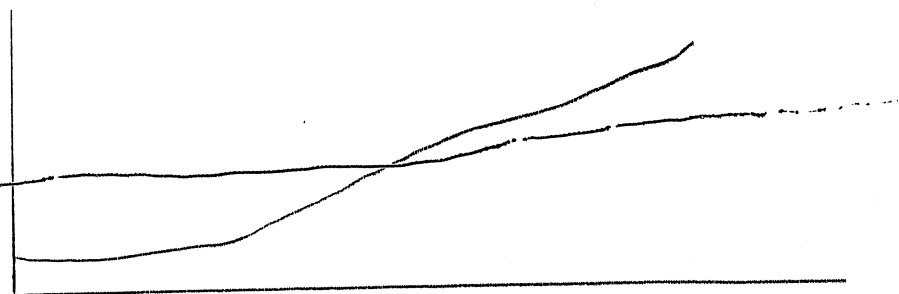


Figure 48. The change in drift speed depending on the value of the gradient pressure.

The authors show that, using these relationships we must bear in mind the errors in determining the angle ψ , which for medium wind speeds reach ± 15 degrees. Thereupon the errors grow with a decrease in wind speed and decrease with its increase. The errors in determining the drift speed reach ± 0.1 kilometers per hour (about 3 centimeters per second).

The studies examined here show the clearly insufficient accuracy in the relationships for computing ice drift. It is entirely apparent that the practical completion of sufficiently accurate ice drift calculations at the present time is possible only for short sections of the course for high wind speeds and only when the wind acts for a comparatively long time not changing substantially its direction or speed, that is, if the ice drift can be considered as a steady motion. On the other hand, however, the errors which are unavoidable in the computation may be inadmissible in practice.

Unfortunately, frequently the need for computing the ice drifts is so great that it becomes necessary, despite the fact that errors may unavoidably creep into the computations, to throw in the probable parts of ice field drifts. In these cases we cannot use the charts of atmospheric pressure averaged for a long period of time (for example for a month), inasmuch as according to the theory studied above and the observations of drift at Severnyy Polyn, the varied directions and speeds of ice drift account for the varied wind speeds.

For this it is necessary first of all to learn carefully the synoptic condition for the entire period for which we are to

compute the ice drift and pick out, if possible, periods when the synoptic conditions represents a steady wind in the area of the probable drift and when the wind pattern was steady. It is necessary to study separately the periods of time when the wind was strong and weak. The computation is best made only for those periods of time when the wind was sufficiently strong (more than 5 - 8 meters per second) and steady in direction. In cases of little or unsteady wind it is better to simply assume a complete absence of drift than to make any computations. The error is expected to be less then, inasmuch as the ice field will move along a complex trajectory in a light wind condition and unsteady direction, hardly making a reciprocal motion. However, to compute all the possible nonlinear motions of ice floes to any degree of accuracy is impossible at the present time.

It is important in computing the ice drift to have in mind a clear picture of the currents in the area where we expect ice drifts, as the speed of the ice drift often can be jointly measured with the speed of nondrift currents, and for this reason all the computations may have nothing in common with actuality.

In computing the drift it is extremely convenient to use the charts of the lines of current, construction of which was described in section on forecasting disturbances. Therefore, after analyzing the synoptic conditions and picking out the period of time with a homogenous wind pattern, it is best to construct averaged charts of the lines of current for them and to compute the drift on the basis of these charts. However, it is necessary first of all to establish from the products of shipboard observations what angle the wind makes

with the direction of the isobars. As was mentioned already, Vasil'ev and Glagolev figured the angle between the wind and the isobars. As was mentioned already, Vasil'ev and Glagolev figured the angle between the wind and the isobars for the Kara Sea area to be about 10 degrees. This value is extremely likely but we cannot use it for other areas as the angle depends on the latitude of the place and on other conditions where were briefly mentioned in the section on forecasting disturbances.

34. FORECASTING THE PERIODS OF SPRING ICE PHENOMENA

If in working out the fall ice phenomena by the basic element on which the periods of the appearance of ice greatly depend we get the heat reserve of the sea, then in working out the spring ice forecasts, that is, the forecast of the opening up and clearing of the sea of ice, the thickness of the ice plays the same part. Actually the greater the thickness of ice the more heat is required to break it up, and consequently, if we suppose that the amount of heat used from year to year is unchanged, then how quickly the ice will break up will depend on its thickness, and so will the periods of opening up and the periods of clearing. Therefore, it is first of all necessary to establish exactly that thickness of ice which should be kept in mind when working out ice forecasts.

V. V. Piotrovich, after intensively studying the observations of ice phenomena on rivers and lakes, contends that the thickness of ice changes within wide limits for a small expanse. Measurements of the ice thickness of a cross-section in the Azov Sea and in other seas showed that the changes in ice thickness at a distance of several meters may reach 40 percent of the mean thickness of ice of a given cross-section. These changes occur for many reasons:

varying periods of ice formation; unevenness of the speed of the currents under the ice; breaking off; piling up; shoving under; etc. In connection with this, and also bearing in mind the difficulty of measuring ice at sea, it is necessary to find a means of determining objectively the thickness of ice without resorting to a great many measurements. The best means, the most correct and objective, would be that proposed by Piotrovich: establishment of the ice thickness from a probability curve constructed on the basis of many measurements from which it was necessary to take the thickness having the greatest frequency. However, we will not use this means at all for the marine conditions in which even one measurement of the thickness of ice is extremely difficult.

Therefore, not having objective data on the ice thickness G. S. Ivanov, E. M. Sauskan and others proposed to determine the thickness of ice formed over the winter months by the sum of negative air temperatures. As we know, the thickness of ice is related to the sum of air temperatures parabollically, however, part of the branch of the curve representing this relationship referring to large thicknesses of ice can be taken as a straight line without much error, which gives us the opportunity of computing the thickness of the ice in the spring time from a linear relationship between the sum of degree-days of frost and the thickness of the ice. Sauskan considers it possible for southern seas (Caspian, Azov, and Black), where a snow cover on the surface is either altogether lacking during most winters or is very slight in spring time when the ice reaches its greatest thickness (not less than 40 centimeters), to use the following relationship between the sum of degree-days of frost and the thickness of ice in centimeters:

$$\Delta \sum (-t) = 18 \Delta h$$

We can also use with great success the general relationship

true for all seas, which is derived from the studies of A. G. Kolesnikov (here it is necessary to determine empirically the constant coefficients for concrete cases). This relationship unlike the one given above is convenient in that it can be applied for any ice thickness since it computes the nonlinearity of the relationship of the thickness of ice to the sum of degree-days of frost:

$$h^2 + bh = a \sum (-t)$$

Using the latter relationship it is easy to compute the thickness of ice for any time. The thickness of ice observed in each separate case will frequently differ from that computed by the indicated means as a result of the uneven growth in area, but the mean values should be close to the computed ones.

Having decided on the method of computation it is necessary to establish the period of time for which we wish to compute the thickness of ice. Obviously this period is that for which the ice thickness reaches its maximum. Then the growth of ice stops and its break-up begins. But determining the maximum thickness of ice is just as difficult from direct observations as it is to determine the thickness of ice at any period of time.

Attempting to avoid this difficulty, Sauskan, from the products of observation in southern seas, determined the relationship between the periods of the steady passage of air temperature through zero degrees in 13 hours and the periods when the thickness of ice reaches the maximum. A steady passage of the air temperature through zero degrees was considered one after which there was no considerable cooling, that is, the sum of the negative temperatures for the

entire period of cooling did not exceed in absolute value the sum of the positive temperatures accumulated before cooling. The relationship of time when the thickness of ice reaches a maximum its thermal conductivity is extremely small because of the great thickness of ice, and the negative air temperatures are not large; therefore, any considerable growth in the thickness of the ice is excluded. At the same time the daily air temperatures become so high that the ice begins to melt. On the surface of the ice we get melted water. At night when the air temperature drops below zero degrees, the melted water becomes coated by a thin crust of ice having the same properties as glass covering a hotbed, that is, not passing long-wave radiation, as a result of which the entire water formed from the ice melting during the daytime again becomes ice at night. Water increasingly forms from day to day on the surface; it fills in all the dents on the ice surface. This melted water which covers a considerable portion of the surface becomes a heat insulator for the ice layer from the falling air temperature taking place at night. All this raises the temperature of the entire ice layer, which also changes its structures. However, when as a result of the drop in daily air temperatures to negative values, the entire melted water becomes ice again, the thickness of ice in its basic mass may again begin growing.

It is apparent that the relationship of the periods of maximum ice thickness to the dates of the passage of 13-hour air temperatures ~~through zero degrees~~ has a meaning only for those areas where the daily air temperature travel is sufficiently great (as, for example, on the Caspian Sea), in the reverse case, however, we must use the relationship of the periods of maximum ice thickness to the periods of the passage of mean daily air temperature through zero degrees.

Just how important a part is played by the ice thickness in forecasting the periods of the opening-up and clearing of the ice is seen from the relationships obtained by Sauskan for five points located on the Caspian, Azov and Black Seas. In these relationships the data of the opening-up or the clearing (D_B , D_O) are determined as a function of the sums of negative air temperatures ($\sum - t_a$), accumulated from the time of the appearance of ice until the passage of the air temperature in 13 hours through zero degrees. In any case we may assume that the earlier the commencement of the opening-up or clearing the earlier will the ice begin to break-up, and the date of the passage of the 13-hour air temperature through zero degrees is an indicator of the time from which the ice began to melt. The named relationships and their characteristics are shown in Table 12.

[See next page for Table 12]

In Table 12 we have the relationships without considering the intensity of the air temperature rise after its passage in 13 hours through zero degrees. From Table 12 we see that 20 percent of the amplitude in many cases makes up more than half the report period, and consequently, the equations shown may be used only in determining the forecasts of the opening up and clearing after the air temperature passes through zero degrees in 13 hours. For a sufficiently good forecast of the passage of the air temperature through zero degrees these relationships could be used in developing long range forecasts also. Table 12 shows also the insufficient reliability of the forecasts for a small report period. Actually, if a forecast could be made for Kherson with an advance report period of

TABLE 12

Name of the points	Equations	General Co- efficient of Correlation	No of years of observations made before the construc- tion of the rela- tionship	Report period in days	Long-range amplitude % days	Reliability Percent
<u>Gur'ev</u>						
Opening up	$D_B = -0.15 \Sigma -t_a + 3.38 D_t + 3.2$	0.90	23	25	20 7	83
Clearing	$D_o = -0.20 \Sigma -t_a + 2.36 D_t + 10.8$	0.95	23	34	20 7	100
<u>Astrakhan'</u>						
Opening up	$D_B = -0.30 \Sigma -t_a + 0.32 D_t - 7.5$	0.85	32	34	20 9	88
Clearing	$D_o = -0.30 \Sigma -t_a + 0.18 D_t + 3.6$	0.81	32	27	20 8	84
<u>Taganrog</u>						
Opening up	$D_a = -0.44 \Sigma -t_a + 0.33 D_t - 7.5$	0.86	28	22	20 9	89
Clearing	$D_o = -0.38 \Sigma -t_a + 0.22 D_t + 7.7$	0.90	19	30	20 9	84
<u>Rostov</u>						
Opening up	$D_o = -0.18 \Sigma -t_a + 0.43 D_t + 7.9$	0.82	47	16	20 9	81
Clearing	$D_o = -0.16 \Sigma -t_a + 0.32 D_t + 17.0$	0.77	47	24	20 9	81
<u>Kherson</u>						
Opening up	$D_B = -0.47 \Sigma -t_a + 0.26 D_t - 13.9$	0.84	24	12	20 8	79
Clearing	$D_o = -0.37 \Sigma -t_a + 0.40 D_t - 6.0$	0.81	25	19	20 8	80

12 days, and the errors stay within the limits of 8 days only in 79 percent of the cases, then the point is that such a forecast can not satisfy greatly all those interested in it. Therefore Sauskan had to study the influence of the intensity of the air temperature rise after it passed through zero degrees for the periods of the opening up and clearing.

If we compute the periods of the full clearing of the sea of ice, then, obviously, the amount of heat necessary for the full break-up of the ice or its complete transformation to water can be determined with the aid of the relationship: $\Delta W = 80 \rho \Delta h$. Here ΔW is the quantity of heat expressed in small calories h is the ice thickness in centimeters; ρ is the density of ice.

Using these relationships and data on heat flow it is easy to compute the number of days which are required for the complete disappearance of ice. Unfortunately, forecasts of the thermal equilibrium or the elements from which it is formed have not been made yet and their calculation from weather forecasts cannot be of the needed accuracy. In connection with this an attempt was made, as in the case of computing the period of the appearance of ice, to substitute the quantity of heat coming from the direct and dispersed radiation, from the heat exchange with the air, from the condensation of water vapor on the ice surface, with the sum of the positive temperatures assuming it as proportional to the influx of heat on the ice surface. For this it is necessary to determine the coefficients of proportionality between the sum of the positive temperatures and the thermal equilibrium. This can be done with the aid of several examples.

For example, Sauskan constructed relationships between the sum of positive air temperatures expressed in degree-days accumulating from the time of the passage of the water temperature through zero degrees in 13 hours to the complete clearing of the sea of ice, and the maximum ice thickness, or, which is the same thing, the ice thickness observed on the day of the passage of the 13-hour air temperature through zero degrees. These relationships were constructed for the same areas of the Caspian, Azov and Black seas (Figure 49). Studying Figure 49 it is easy to see that the slope angle of the lines of regression to the abscissa in all five cases was the same. Conversely, the section on the ordinate which will be formed if we extend the lines of regression to intersect it and included between the origin of the coordinates and the point of intersection was essentially different. This shows that the general equation for the five points under study can be written in the form:

$$\sum t_a = 1.4 h_{max} - c$$

In this equality the coefficient of proportionality, 1.4, representing the slope angle of the lines of regression to the abscissa, remains unchanged for all five relationships at the same time that c changes from point to point. It is easy to see that c increases from west to east. For Cur'ev this quantity reaches the maximum value. Probably, the change in c can be explained by the decrease in the weight of the direct solar radiation in the general thermal equilibrium, because of the decrease in the continental effects of climate from east to west. Together with this, the daily travel of the temperature also decreases, and, consequently, the time between the passage of the mean daily temperature through zero degrees and the temperature observed in 13-hours, that is, at the

time close to when the greatest amount of heat is coming from the sun.

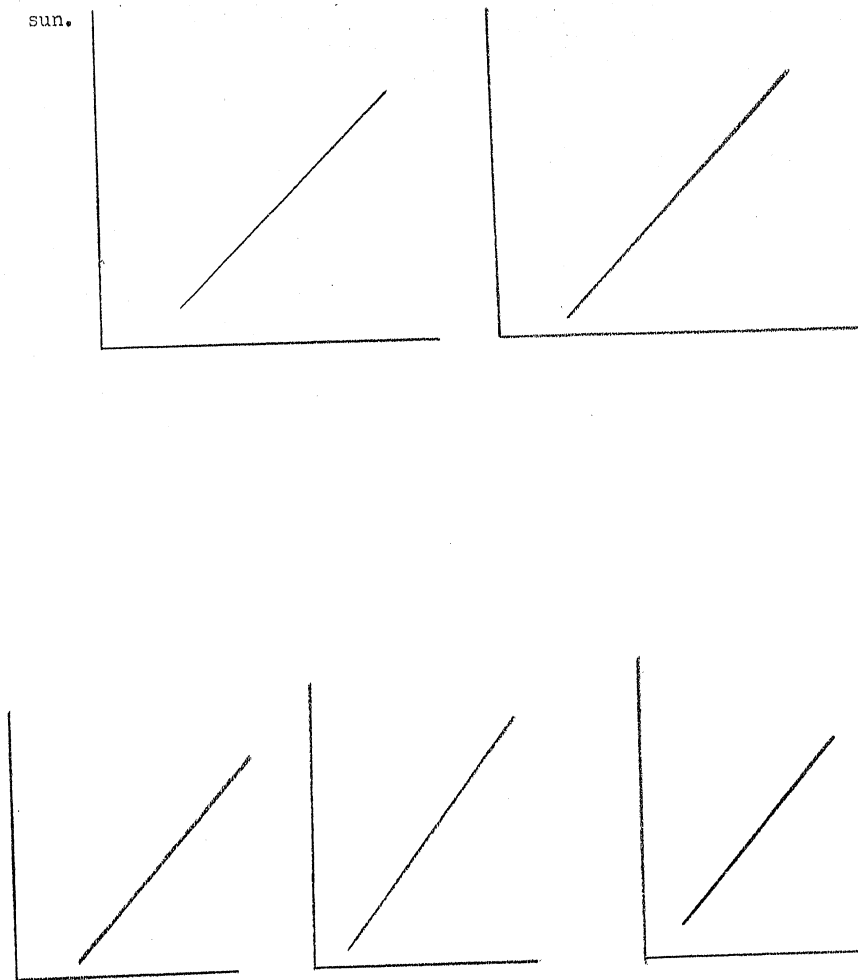


Figure 49. A relationship of the sums of the positive air temperatures to the maximum ice thickness: a - Gur'ev, 6 - Astrakhan', p - Taganrog, z - Rostov, d - Kherson.

If we are to compute the increase in the sums of the posi-

tive temperatures from day to day and the melting of the ice related thereto, then the relationship being studied may be rewritten in the form:

$$\Delta \Sigma + t_a = 1.4 \Delta h$$

Now it is easy to determine how the thermal equilibrium of the ice surface is related to the sum of positive temperatures. For this purpose it is enough to compare with one another the equations:

$$\Delta w = 80\rho \Delta h; \quad \Delta \Sigma + t_a = 1.4 \Delta h$$

From these two equations it is easy to establish a relationship of the thermal equilibrium to the sum of the positive temperatures. This relationship can be written in the form:

$$\Delta w = 53 \Delta \Sigma + t_a,$$

if $\rho = 0.92$.

As we will show below the relationship of the thermal equilibrium to the sum of positive temperatures derived by E. M. Sauskan and described above were almost identical with the relationship obtained by other authors.

Above we determined the relationship linking the maximum ice thickness with the sum of degree-days of frost accumulated from the time of the appearance of ice to the time when the ice reaches maximum thickness. This relationship can be converted to another form if Δh in this relationship is replaced in the equation:

$$\Delta w = 80\rho \Delta h$$

Then it is easy to determine the amount of heat which was lost through the surface of the ice cover in the form:

$$\Delta w = 4.10 \Delta \Sigma - t_a$$

Obviously the amount of heat lost in the formation of

ice must be equal to the amount of heat which will be required in breaking up the same volume of ice, and as a result of this we can equate the right sides of the equations determining the relationship of the sum of positive and negative temperatures with the thermal equilibrium of the ice surface:

$$\Delta \Sigma + t_a = 0.08 \Delta \Sigma - t_n$$

This equation allows us, disregarding the changes in ice thickness, by simple computations to find from the sum of degree-days of frost accumulated during a winter the sum of degree-days of heat necessary for a complete break-up of ice. That condition, that the sum of degree-days of frost required in forming the ice corresponding to the thickness is 12.5 times greater than the sum of degree-days of heat used in its break-up, is explained by the low thermal conductivity of the ice which plays a large part in the formation of ice (the ice melts on top, but forms on the bottom). Knowing the amount of degree-days of heat necessary for the break-up of the ice, and the forecasts of the path of the air temperature, it is easy to compute the number of days during which the process of break-up of the ice cover will take place, and on the basis of this computation to forecast when the sea will be cleared of ice. It is apparent, that the computation of the sum of negative and the sum of positive temperatures must be made from the observations of the station for whose area the forecast is made.

The examined means of computing the time when the sea will be cleared of ice is significant only for the points located on the Black Azov and Caspian seas, and then for only those cases when the maximum ice thickness varies within the limits indicated above.

To apply this means of computing in other areas it is necessary to establish a new all the coefficients of proportionality.

Actually, if we were studying the ice cover whose surface is covered by a thick layer of snow (in southern seas, as a rule, there is very little snow), the relationships between the sum of degree days of frost and the sum of degree-days of heat would have to change substantially. A more general relationship used in computing the sum of degree-days of frost is the equation mentioned above:

$$h^2 + bh = a \sum -t$$

in which it is also necessary to determine the coefficients for various climatic conditions. The relationship of the sum of degree-days of heat to the thermal equilibrium cannot remain constant for all conditions of climate, and consequently, nor can the sum of degree-days of heat to the thickness of the ice cover. New coefficients can be established by the method described above, but this problem can be solved by another method.

For example, P. P. Kuz'min and P. P. Nikiforov built relationships between the thermal equilibrium of the ice surface and the positive air temperatures. The relationship obtained by Nikiforov is shown in Figure 50. The air temperature taken for this was the mean daily one. From Figure 50 it is easy to note that sometimes the thermal equilibrium may be negative for positive air temperatures, and at the same time the thermal equilibrium (as is noted in Kus'min's work) can be positive for negative temperatures. This explains the melting of ice noted above for negative mean daily temperatures. From here it follows that a sufficiently accurate computation of the melting of the ice is possible only in computing the complete thermal equilibrium. But it has been said

many times that it is impossible at the present time to predict with sufficient accuracy the elements of thermal equilibrium. Therefore, it is necessary to use the relationship of the thermal equilibrium of an ice surface to the positive air temperatures. As we see in Figure 50 it is so close that it allows us to use this method. With some approximation the relationship built from the products of observations made at the Azov and Taganrog stations located on the Azov Sea may be expressed by the equation:

$$w = 50t_a$$

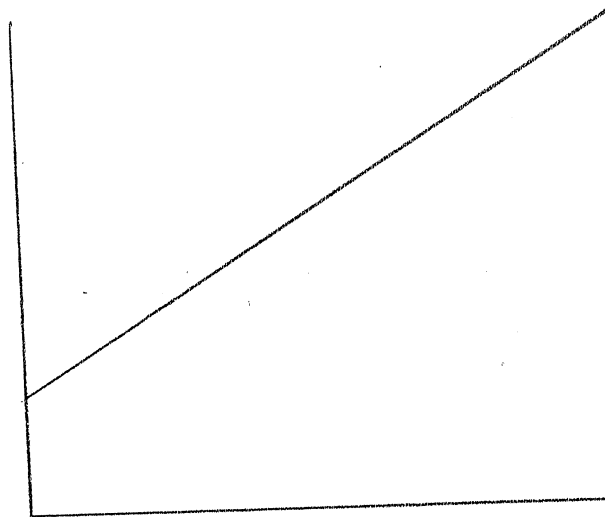


Figure 50. The relationship of the daily equilibrium of the ice surface to the mean daily air temperature.

Kuz'min got similar relationships from the observations of the melting of glacier ice:

$$w = 51 t_a$$

As was indicated above, Sauskan also got almost the same coefficient:

$$w = 53t_a$$

These relationships were obtained for the melting of almost snowless ice in areas located approximately on the same latitudes. Kuz'min undertook a similar job of studying the melting of ice in the Volga, Don, Vologda, Sukhona and Vaga Rivers analyzing the products of observations for 29 points. As a result for higher latitudes and considerable snow coverage on the ice the relationship of the thermal equilibrium to the air temperature was substantially different from those studied above:

$$w = 76t_a$$

The recomputation of the amount of heat, calculated with this relationship, on the ice thickness melting for one degree of positive temperature shows that one centimeter melts for one degree. In the case of the relationship obtained for southern seas, however, one centimeter melts for 0.7 degrees. It is possible, as Kuz'min supposes, that the too large coefficient 76 was obtained as a result of the melting in the rivers taking place from ^{the} top surface of the ice and from the lower surface, inasmuch as in the spring large amounts of melted waters having a temperature of over zero degrees and ground waters also with high temperatures come from the rivers, and in analyzing the products of observations all the melted ice was considered as having melted from the upper surface. Using the relationship of the thickness of the melted ice to the sum of positive air temperatures it is easy to compute the number of days required for the air temperature necessary for the melting of ice

with a given thickness to reach that value and as a result to have a complete clearance of the sea of ice.

The problem of deciding the question of the time of the opening up of the sea is somewhat more complex. If the sea is cleared of ice as a result of its melting, then the opening up occurs when the ice still has a rather considerable thickness. It is well-known that the opening up of the ice cover does not take place for a uniform thickness of ice in several areas of even one sea. It is apparent that an especially great difference in the ice thickness for which its break-up occurs is observed in comparing the given observations of the opening up with various seas. The ice thickness in opening up is not uniform, even at the end the same point observed in various years. The only regularity which is observed in this phenomenon conforms to a simple rule: the greater the maximum ice thickness during a winter the bigger the thickness of ice for which an opening up will take place. This relationship is clearly explained by the fact that when the ice melts on top an increase in the temperature of the ice throughout the entire thickness takes place gradually. At some moment of time the entire thickness of ice will have the same temperature -- the melting temperature of ice. As a result of this there will be a change in the ice structure: it will become less solid and as a result it will begin to break-up mechanically. But always parallel with the increase in ice temperature, especially in sea ice formed from salt water and therefore having many cells filled with brine, we get a gradual oozing of the ~~water accumulating~~ on the surface from the melting. The oozing takes place through the microscopic pores or cells filled with brine. In this manner together with an increase in the temperature of the ice and the decrease of

its solidity connected with it, we observe a decrease in the solidity because of gradual erosion by the melted water.

To determine the ice thickness observed in the opening up process we build for each from the maximum thickness for the winter a relationship of the type:

$$h_{\text{opening}} = kh_{\text{max}}$$

From this relationship we compute the ice thickness at which the opening up must take place for a given year. Further, we find the difference between the maximum ice thickness and the ice thickness at which an opening up must take place according to the computations. From the relationship of the sum of positive air temperatures to the thermal equilibrium of the surface of melting ice we compute what sum of positive air temperatures is necessary for melting the ice thickness equal to the difference between the maximum and the opening up thicknesses. Having determined the sum from the forecast of mean daily air temperatures we find the number of days required for the accumulation of this sum, and compute from the number of days the date when the opening up will occur.

It is important to bear in mind the incompleteness of the computations described above. We can, therefore, approximately compute only the thermal break up of the ice cover. At sea quite frequently the break-up of the ice occurs for mechanical reasons also. Even a solid winter land floe can often be torn loose and carried out to the open sea by a storm that has come up. Understandingly, in the spring when the ice is considerably weakened by a thermal break-up it is even more subject to mechanical reaction.

In many cases in seas with flood tides we can notice that

the opening up of the sea in this or that area is most frequently observed during syzygial flood tides when the amplitude of variations in the sea level is especially great. It may be assumed that the sgonno-nagonnye variations in the sea level may also be an essential influence on the time of the opening up of the land floe. Unfortunately, there have been no efforts made which would have allowed us to quantitatively evaluate the influence of the mechanical factors on the periods of opening up. It is necessary to calculate these factors by a careful analysis of the products of observations during the time of opening up.

As the first step for each separate area we analyze the influence of the variations in sea level on the periods of opening up. Next we study the influence of the wind on the same periods. A careful study of the products of observations allows us to introduce corrections tentatively into the forecasts of the periods of opening up, developed by computing only the thermal break-up of the ice.

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CHAPTER VII

THE CONCEPT OF LONG-RANGE FORECASTS35. SOME INFORMATION ABOUT LONG-RANGE HYDROMETEOROLOGICAL
FORECASTING

The basic information about short-term hydrometeorological forecasting studied above shows that for short periods of time the compiling of forecasts of the sea condition (in a wide sense of this word) is difficult. In many cases the phenomena are as yet little studied sometimes as a result of the inaccessibility of the products of observations and sometimes because of their complexity. Therefore, the relationship used in computations do not yet have the accuracy necessary for full satisfaction of inquiries from organizations interested in forecasts.

An overwhelming majority of the forecasts of the separate elements of the sea pattern may be built only on the basis of a weather forecast or its separate elements. The necessity of building forecasts of the sea pattern on weather forecasts is the weakest spot in the development of short term forecasts of the hydrometeorological sea pattern, inasmuch as predicting the weather is even now one of the most difficult problems of modern science. Short-term weather forecasts, despite the big strides toward solving this problem remain insufficiently accurate and have insufficient justification. Even more complex is the matter of the long-range weather forecasts. The accuracy and justification of long range weather forecasts is considerably less than for short-term ones and is limited by natural reliability.

If the long-range weather forecasts had a sufficiently high

justification, at least as high as that of the short term one, then it would be possible to apply all the methods presented above in forecasting the sea condition. The existing justification of long-range weather forecasts is such, that to apply for long-range forecasts the method described above in the section on short-term forecasts of the sea pattern is entirely inexpedient.

That is why a necessity arises of developing special methods for long-range forecasts of the changes in various elements in the sea pattern most important for navigation.

The most simple method of forecasting is the purely statistical method. It boils down to this, that in forecasting we should always repeat the same value of the predicted element having a high probability in comparison with all the other values of this element.

Frequently this value of the element of the pattern having a high probability coincides with the mean value of the predicted element, but often we may note cases of noncoincidence of the mean value with the probable. To determine accurately the probable value we resort to constructing distribution curves, taking from them the value giving the greatest frequency. Quite frequently the repeating value may be determined by subtracting the double mean value from the triple mean (median) value of the examined element. The mean (median) value denotes the one value of the series arranged in ascending order of the numbers of several distributions which bisects the series into two equal parts, that is, into two such parts in which ~~the~~ above and below the mean value there are the same number of terms.

But statistical forecasts cannot be too interesting, inasmuch

as by this method it is never possible to predict practically the especially important greatest or least values in the travel of any element of the sea pattern. Therefore, there is no need to dwell on an accurate exposition of this method here.

In the travel of the elements of the pattern we can frequently notice periodic variations. The periodic variations are expressed mathematically by periodic functions. If a phenomenon has periodic characteristics then we must fill in the equation:

$$f(x) = f(x + \tau)$$

This equation implies that on the expiration of the period of time τ the function $f(x)$ has the same value as for the time x . To determine the function $f(x)$ for any value of the argument it is enough to know it in the interval from x to $x + \tau$, where τ is the period of the given function. As an example of a periodic function we can consider the trigonometric functions:

$$f(x) = \sin\left(\frac{2\pi x}{\tau} - \varphi\right).$$

Actually it is easy to see that with the expiration of time $t + \tau$, $t + 2\tau$ or $t + n\tau$ the function acquires the same values. Examples of periodic variations in the travel of hydrometeorological elements can be pointed out frequently. For example, the daily travel of water temperature or air temperature, the yearly travel of the water and air temperature and so forth. Actually, it is well known to everyone that in January in the northern hemisphere, at any point located beyond the limits of the torrid zone, it is always cold at the time when in July it is always warm. In this manner the periodic variations in the air temperature allowed us firmly to set the rule,

that in the winter we should expect the cold and in the summer the heat. Similarly, another rule is well set forth, that it is warmer in the daytime than at night, regardless of the time of year. However, a more detailed familiarity with the travel of the air temperature shows, that it is not equally cold every January, just as it is not equally warm each July. If we compare the mean monthly air temperatures for many years but for the same month, then it is easy to be convinced that the temperature variations from year to year are very considerable, sometimes reaching 10 degrees and more.

The reason for such temperature changes -- , at first glance resembling periodic variations, but after a more careful familiarization they are not purely periodic and do not conform to a strictly mathematical rule of repeating the exact value of a function with the expiration of the period -- is hidden in the purely periodic entrance of heat from the sun causing the air temperature to rise in the summer and drop in the winter, and in the imposition on this purely periodic phenomenon of some other nonperiodic influence causing periodic temperature variations. It is obvious that if the temperature variations had a purely periodic characteristic then the temperature forecast would not present any difficulty. It would be enough to use one year in order to compute the temperature for the following year for any number of years ahead. The simplicity of setting up forecasts in cases of periodic variations is one of the reasons for the many attempts by scholars to find the correct periodic or at least cyclic variations in the travel of hydrometeorological and hydrological elements of the pattern. Cyclic variations are those for which the periodicity first appears in the travel of any element then disappears (dies down), after which the length of the period varies around some

mean value and is not strictly constant as in the case of the periodic variation.

For example, the many attempts, by numerous authors working in the most varied regions, at correlating the cyclic variations observed in the formation of sun spots with the changes in the travel of hydrological or meteorological elements of the field are well known. The spot forming action of the sun has a sharp cyclic nature, for which the mean period of the cyclic action of the sun is close to 11 years. Quite reliably we have determined the relationship between the disturbance of the earth's magnetic field and the sun spots. There are many studies indicating the good relationship in the travel of the air temperature in areas removed from the equator to the sun spots. The German scholar Bauer contended that in the travel of the air temperature in the western European area and in the solar action there also is a definite relationship. The Soviet scholar V. Yu. Vize contended that in Lake Victoria, located in Africa, the level varies depending on the changes in the number of sun spots. He also found the changes in the iciness of the arctic sea with a period equal to half the period of the variations in the number of sun spots, etc. A count of the authors and the elements with which the numbered sun spots were connected could be considerably lengthened; however, there is no necessity in this, inasmuch as most of the relationship ever obtained of the solar action, after many observations, did not stand up. As a result, despite the volume of work undertaken previously in this direction, the question of the relationship of the travel of the elements of the hydrometeorological pattern to solar activity up to the present remains open. Moreover, in the past few years scholars have been attributing less significance

to such types of relationships because empirically they are not confirmed by an increased number of observations, and there is no reasonably satisfactory theory correlating the influence of sun spots on the processes arising in the earth's atmosphere.

There were also many attempts at discovering in the travel of the processes arising in the atmosphere and the hydrosphere other periodicities not related to solar activity. In this it was assumed that the path of this or that element of the pattern is composed of several simple periodic variations which in sum are extremely complex. Therefore, the discovery of disguised periods was made a problem of research. But it is necessary to note here the exceptional difficulties with which the researcher comes up against in attempting to solve this problem. At the present time there is no method allowing us to solve precisely the problem of the reality of the periods which can be obtained as a result of one or the other analysis. Neither is there a method which would help in the objective and correct solution of the problem of, from which simple variation the more complex one is made if the periods of these simple variations are unknown. Conversely, the solution of the problem is very simple if the periods of the simple variations from which the more complex ones are compiled are known, but other elements of the variations are unknown as, for example, amplitudes and phases. Maybe it is due to the complexity of this problem, or more correctly, due to the absence of real, strictly periodic variations, that not one of such studies ended in undisputable success. Up to the present time there is no method based on the periodicity of the travel of any element of the hydrometeorological pattern which would allow us to develop forecasts with an accuracy sufficient for practical pur-

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poses.

Frequently in developing long-ranged forecasts we use the method of analogies. The substance of this method consists in that, using the products of observations for many years, we can compare the travel of hydrometeorological processes observed in the current year. Sometimes we can select a year in which the travel of the processes is rendered in many similar travels of processes in the current year. In this case one reasons in the following manner: if in the period of several months the processes of the current year were similar to processes of a particular year, then we can assume that in future during a certain period of time the processes will remain similar; consequently, we can take the observations of a similar year and use them as a basis of forecasts of a later time in the current year. This method of forecasting is usually used when there is no better method and there is a great need for developing a forecast of some element.

As a rule such a forecasting method, not strengthened by some more well-grounded development, gives unsatisfactory results for the following reason: complete similarity in the travel of the elements of the pattern in the current year and in any of the previous years never occurs, and an uncomplete relationship cannot produce a similar travel of the processes in the future -- if it can then only accidentally. Here it is necessary to add that the analogy is usually made by sight, approximately, inasmuch as for the quantitative determination of the degree of similarity there would not be enough time for the forecaster, and there would have to be special methods set up for determining the degree of similarity. Hence, the method of

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selecting similarities in no case can be recommended for practical application in developing long-term forecasts.

Many years experience in developing every possible method for long-term forecasts of the elements of the hydrometeorological pattern shows that a method deserving some attention can be obtained only when it has a physical basis. On the other hand most of the methods of long-term forecasting, after some experience in applying them in practice, in formulating forecasts, are entirely unsatisfactory. Below, as examples of particular methods for long-term forecasts deserving attention, are presented some developments in ice and sea level forecasts.

36/ LONG-RANGE FORECASTING METHODS FOR ICE CLEARING IN THE WHITE SEA

As was already noted in the description of the short-term forecasting method for spring ice phenomena, the time when the sea will be cleared of ice is greatly dependent on the ice mass formed for the winter. There the maximum ice thickness or the sum of degree-days of frost for a winter period served as indicators of ice mass. It is quite obvious that we cannot ignore the ice mass formed at sea during the winter even in the case of long-term forecasting. The more ice at sea, the greater the time required for its melting and break-up. V. V. Timonov and K. I. Kudryavaya adopted the ice density for the characterization of an ice mass, inasmuch as for the White Sea where forecasting ice is found and land floes are frequently torn loose by the wind, the ice thickness could not serve as the measure of iciness. It is extremely probable that the sum of

degree-days of frost for a winter also could serve as a good and maybe as a better indicator of the iciness of a sea than the density of the ice. The trouble is that the ice density observed at shore stations cannot serve always as a measure of the iciness of an open sea and, besides that, it cannot measure sufficiently reliably the ice mass, inasmuch as the thickness of floating ice whose density is being determined, is not measured. It can happen that for a large density there will be a small ice thickness, and, conversely, ice of little density may have a large thickness. However, using the sum of degree-days of frost as a measure of the ice mass in the White Sea is necessary in a special study, since for the White Sea, the carrying away of the ice, which takes place during wintertime is characteristic. But, unfortunately, the carrying away of the ice from the White Sea is quantitatively still insufficiently studied; to talk of the possibility of replacing the ice density with the sum of degree-days of frost, therefore, is still premature. Besides this, the ice density in the work of Timonov and Kudryabaya before its use in building prognostic relationships went through an extremely careful critical analysis, and therefore its use in prognostic relationships should not be doubted.

Just how essential the influence of iciness on the date of ice clearing for a sea is, can be seen from the coefficients of correlation obtained by the authors of this work. In correlating it became clear just how much the iciness of a succeeding month depends on the preceding one or how great the ice inertia is:

Months	II-I	III-II	IV-III	V-IV
Coefficients of Correlation	0.58	0.36	0.80	0.80

Let us apply the coefficients of correlation representative of the abnormality of the periods of ice clearing in relation to abnormality of the iciness of different months:

The months for which the iciness was correlated with abnormal ice

clearing periods	I	II	III	IV
The coefficients of correlation	0.41	0.61	0.52	0.86

As a measure of the intensity of heat influx during the spring the mean monthly air temperatures ^{air} was used in this work. The influence of the air temperatures in each separate month on the intensity of the spring processes of ice clearing can be seen from the following data:

The Months for which the abnormalities in the temperature were correlated with the abnormalities in the periods of

clearing	II	III	IV	V
The coefficients of correlation	0.48	0.59	0.68	0.79

A comparison of the correlation coefficients in the last two tables for the same month shows that the influence of temperature and iciness on the periods of ice clearing has the same order, and even the coefficients in the relationship of clearing periods and iciness abnormalities slightly exceed the coefficients in the relationships of clearing period and temperature abnormalities.

Besides this, to measure the peculiarities of atmospheric processes of each year Timonov and Kudryavaya used the air pressure gradients over the White Sea, computed from the chart of the

mean monthly pressure. The pressure gradients were determined for a distance of 100 kilometers (for 500 kilometers from the center of the sea).

A comparison of the abnormalities in clearing periods with the transfer of air masses in March computed from the pressure gradient in the direction of the meridian and in the direction of the parallel gave coefficients of correlation indicating that the transfer of air masses from south to north or, conversely, from north to south plays a basic determining part in the development of spring ice processes in the White Sea, a secondary part being played by the transfer of air masses along the parallel.

The coefficients of correlation representing this relationship were the following:

$$\Delta D = f(V_{III}^{NS}), \quad r = 0.64$$

$$\Delta D = f(V_{III}^{NS}), \quad r = 0.15$$

The coefficients of correlation in the case

$$\Delta D = f(V_{II}^{NS}, V_{III}^{We})$$

were equal to 0.67.

The maximum of iciness in the White Sea is observed in March, and, as will be recalled, the coefficient of correlation between the abnormalities in the iciness in March and the abnormalities in the ice clearance periods of the White Sea is sufficiently high (0.52). At the same time April is the first spring month in which there is a noticeable increase in the air temperature and a decrease in the

iciness. The coefficient of correlation representing the relationship of the abnormality in the April temperature to the abnormalities in the ice clearing periods of the White Sea, is also sufficiently high (0.68). Bearing in mind that the clearing of the White Sea takes place in May-June and only in some years at the end of April, the possibility of using the relationship of March iciness abnormality (ΔI_{III}) and April air temperature (Δt_{IV}) to the abnormalities in the clearing periods (ΔD). With this aim a relationship of the following type was established: (

$$\Delta D = A \Delta I_{III} + B \Delta t_{IV}$$

where A and B are constant coefficients.

The general coefficient of correlation representing this relationship was equal to 0.76, and the probability of the error not exceeding 20 percent of the long-term amplitude was 85 percent.

The month during which the break-up of the basic ice mass occurs is May. It is easy to be convinced of this if we use the May temperature abnormality (Δt_V) as a third variable in the previous relationship, that is, construct a relationship of the following form:

$$\Delta D = A_1 \Delta I_{III} + B_1 \Delta t_{IV} + C_1 \Delta t_V$$

in which A, B and C are constant coefficients. The coefficient of the majority correlation increases to 0.94, and the reliability to 95 percent.

In this form the equation loses its prognostic meaning, as an equation designated for forecasting ice phenomena independent

of the weather forecasts. In order to use this equation it is necessary to use the forecast of the mean temperature in May.

As was already indicated, the relationship between the ice clearing periods of the White Sea and the transfer of ice masses over the sea in March was established by the authors of the method. And they used this relationship in building the prognostic equation independent of the necessity of using the forecast of mean monthly air temperature for May. As a result several relationships were built of the type

$$\Delta D = A_2 \Delta V_{III}^{WF} + B_2 \Delta V_{III}^{NS} + C_2 \Delta t_{III} ;$$

$$\Delta D = A_3 \Delta t_{IV} + B_3 \Delta I_{IV} + C_3 \Delta V_{III}^{NS} .$$

The first of these relationships has the coefficient of correlation $R = 0.82$ and a 95 percent reliability, in the second $R = 0.94$ and the reliability 95 percent.

The relationship obtained in this manner can be used in forecasting the ice clearing periods of the White Sea. However, this method is not without some shortcomings. The greatest of these is the following. These relationships were built for a comparatively short time -- 20-25 years. In increasing the number of observations we must bear in mind some possible decrease in the correlation coefficients and the reliability. Besides, as was indicated above, the density of the ice taken as a measure of the iciness of the White Sea before its use in building the relationships underwent a careful critical analysis. In making up operational forecasts, it is obvious that such an analysis will be entirely impossible.

In this case each individual station will have to decide on the ice density in the area by the telegraphic data, that is, by data which is known to have a certain percentage of errors. Besides that, all the relationships studied above were constructed for data averaged from several stations at which ice clearing of the sea originates; the ice density also was averaged for several stations, and the same steps were taken in computing the abnormality of the temperature.

It is obvious that in using these relationships in operational work difficulties may arise. For example, if for some reason the information from some stations is temporarily or entirely missing, then difficulties will appear in computing the mean values of the ice density and air temperature.

37. THE METHOD FOR LONG-RANGE FORECASTS OF AUTUMN ICE PHENOMENA

As an example of the methods used in long-range forecasting of autumn ice phenomena we can use the method proposed by Yu. V. Istoshin for the northern part of Tatarsk Strait and the estuary of the Amur River. The basis of his method is the consideration of the peculiarities in monsoon weather pattern, which is quite clearly outlined in the Far East. The monsoon phenomena in relation to the theory of monsoons developed by V. V. Shyleykin can be schematically represented in the following manner.

Oceans and deep-water seas during a summer accumulate very large heat reserves thanks to the large thermal capacity of water and the mixing of waters, but heat up little in comparison with the continent. The temperature of the ocean surface in the summer is less than the temperature of the continent. Continents composed of moun-

tain rock having a lower thermal capacity and a small thermal conductivity in comparison to the ocean waters (bearing in mind the turbulent thermal conductivity) are heated up in summertime to a rather higher temperature than the ocean and sea. In the cold part of the year the ocean, having accumulated a large amount of heat, cools off considerably more slowly than the continents; therefore, the surface of the ocean is warmer than the surface of the continents which quickly expend the heat accumulated in the summertime as soon as the influx of solar heat becomes less than the amount of heat expended in radiation. In this manner in the cold part of the year the sea surface is warmer than the surface of continents, and in the warm part of the year, conversely, the surface of the continents is warmer than that of the ocean. In the cold part of the year the air cools over the continent and warms up over the ocean. In the warm part of the year the air heats up over the continent and cools over the ocean.

In this manner the air in the cold and warm part of the year heats up unevenly over the continent and the ocean. In the cold part of the year the air over the sea heated to rather higher temperatures than over the continent, has a lower density and at some altitude over the sea begins moving towards the continent over which the air becomes more dense as a result of cooling and, consequently, more heavy. In connection with this the heavier air from the continent in the surface layer begins to move in the direction of the ocean. Arriving at the ocean and moving along over its surface it heats up gradually, becomes lighter and, continuing its horizontal motion over the ocean surface rises up simultaneously to higher layers where it gradually reverses its direction of motion, that is, begins to move again in the direction of the continent. Over the

continent it again cools, becomes more dense and simultaneously loses altitude gradually coming down to the land surface.

In this manner the process continues until the temperature of the ocean surface is higher than the temperature of the surface of the continent. In the spring when the temperature of the ocean surface and the continent become more equal, the motion of the air caused by this temperature difference stops and starts up again only when the temperature difference in the surface of the ocean and the continent arises. But in the summer this difference has the opposite sign. The continent becomes warmer than the ocean. In this case ⁱⁿ the lower layer the air moves from the ocean towards the continent, and in the upper from the continent towards the ocean. Usually, when we speak of the monsoon wind we have in mind the wind acting in the lower layer, that is, moving in the summer from ocean towards continent, and in the winter from continent towards ocean, and we forget the existence of opposite or, as they were called by V. V. Shuleykin, antimonsoon air flows due to whose existence the system of monsoon winds is possible.

It is necessary to add that the intensity of the monsoon and antimonsoon flows depends on the difference in temperatures of the ocean and the continent. The greater this difference the greater the speed of the monsoon and antimonsoon flows. Inasmuch as the biggest difference in the temperature of the ocean surface from the temperature of the ocean surface from the temperature of the surface of the continent is reached in the middle of the winter, that is, approximately in January, the greatest force and frequency of monsoon winds of the winter type are reached in January. Conversely, the summer

monsoon gets its development in July. In the spring and fall, however, before the change of the monsoon, the monsoon type winds have the least frequency and the lowest speed. Besides that, the earlier the change in sign of the difference in temperatures between the ocean surface and the surface of the continent takes place, the earlier the change in winds from the summer type to the winter takes place also or, conversely, from the winter type to the summer. One of the main reasons for the change in sign of the difference in temperatures of the surfaces of the ocean and continent is the temperature of the ocean waters.

As was indicated, the continent expends heat very quickly and the ocean slowly. Therefore, the surface of the continent heats up with ⁱⁿ approximately the same time to the greatest temperatures and cools to the lowest. The heating up and the cooling occurs in approximately the same periods to temperatures for which usually the sign in the difference in temperatures of the ocean and continental surface changes. The ocean has considerably more thermal inertia. If we assume that the ocean surface is heated more for some reason during a summer in any specific year than it would be usually, then the difference between the temperatures of the continental surface and the ocean surface will change signs from summer to winter earlier than in usual periods. This will occur because the continent cools as quickly as in normal years and the temperature of the continent evens out with the ocean temperature earlier. Since the ocean in a given year has a higher temperature than usual. The earlier change in the sign of the temperature unavoidably will cause an earlier start of the winter type monsoon winds. Besides this, continuing to cool with the same speed, the continent will have a

greater and greater difference in temperatures with the ocean, and large temperature difference unavoidably must cause an increase in the monsoon flows. Consequently, the air masses cooled on the continent will travel to the ocean surface more quickly. As a result the air temperature in its transfer from the continent to the ocean becomes lower than in a normal year. Autumn and the beginning of winter in this case will be earlier at the shores of the sea and colder than usual.

The variation in the water temperature of the ocean is not the only cause altering the time of the change in the sign of the difference in continental and ocean surface temperatures. Monsoon processes are not the only things that develop in the atmosphere. The change by the air occurs also as a result of the essential difference in the temperatures of the air moving in northern and southern latitudes. The motion of the air is always, or, more precisely, in most cases, determined by two thermal flows: monsoon and interlatitude. It is easy to imagine a case when for some reason the interlatitude flow of heat directed from south to north, whose path goes across the continent, was more intensive for a given year. Under these circumstances a slow cooling of the continent takes place and later a change of the summer monsoon to the winter type. If the intensified flow of warm air masses from southern latitudes to the continent does not stop even after the ~~change in the monsoon~~, then obviously the difference in the temperatures of the continent and the ocean will be less than normal, the winter monsoon will be less intensive, and the fall and beginning of winter later and warmer. If the path of cold air moving from north to south is headed across the continent, then the change in the monsoon and the intensity of

the monsoon will be greater than for years normal in their temperature conditions.

Above we studied a case when the water temperature in the ocean was higher than normal, as a result of which there was an aggravation of the winter monsoon and a severe fall and winter. The reverse situation is also true: if the ocean temperature in any year up to the beginning of the fall processes is less than normal, then we may expect a later and milder fall and beginning of winter. These are the peculiarities of the monsoon pattern used by Yu. V. Istoshin in developing a procedure of forecasting fall ice phenomena for the northern part of the Tatarsk Strait and the estuary of the Amur River. In describing the peculiarities of the monsoon weather pattern there is no indication of how the temperature of the ocean surface and the continental surface is computed, inasmuch as the discussion may be about the temperature of some specific area of the ocean surface or continent about the mean temperature of the ocean or continental surface, about the mean temperature of some areas of the ocean or continental surface etc. Actually though, the boundaries to which the activity of the monsoons extends have not yet been exactly determined, either on the ocean side, or on the continents side. Even though it is entirely apparent that this problem plays an extremely essential part in the selection of an area for computing the temperature suitable for characterizing the monsoon pattern, Istoshin avoids the solution of this problem. He does not even make an attempt to solve it in view of the lack of necessary information. The temperature of the warm Tsushima current is taken as an index of the temperature of the ocean surface. This selection to a considerable extent is dependent on having the necessary products of

observations for many years just for this area and we cannot admit it to be unsuccessful. The Tsushima current is one of the branches of the warm Pacific Ocean Kuro-Sivo current originating from the Northern Equatorial Current. We can assume that the pattern of the Tsushima current is closely related to the activity of the Northern Equatorial Current, the Kuro-Sivo current and with the extension of Kuro-Sivo which after breaking away from the Tsushima current, moves along the eastern shores of Japan and thereafter turns to the north-east. This relationship has not yet been proved, but its existence may be considered entirely probable.

Istoshin used the mean water temperature on a cross-section through the Korean Straits in which the depth reaches 160 meters as a measure of thermal condition of the Tsushima current. He took not the surface temperature, but the mean temperature for the entire cross-section from surface to the bottom. The mean temperatures on this cross-section were computed for each month separately from May to August, and then totaled for this entire period of time. Obviously the mean temperature of the cross-section is more indicative than the surface temperature. If the surface temperature of the water may not prove to be indicative of the thermal content of the Tsushima current, then the mean temperature of the cross-section is the value, sufficiently steady, which is evidenced by the relationship of the sum of mean temperatures of the water in The Korean Strait from the surface to the bottom in the period from May to August to the mean water temperature on the same cross-section in October (Figure 51). This relationship shows that the sum of the water temperatures taken on this cross-section for the summer months

from May to August can replace the water temperature on the same cross-section in the autumn period. This circumstance is very important in increasing the report period of forecasts.

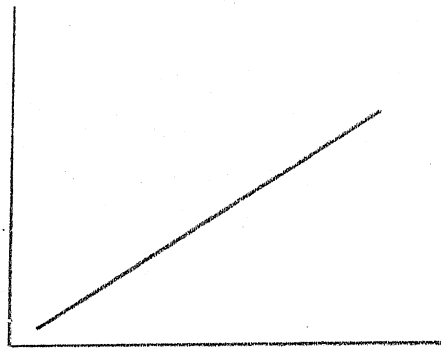


Figure 51. The relationship of the mean monthly water temperature in October in the Korean Strait to the heat content for May-August.

It appeared that the sum of mean water temperatures in the Korean Strait taken for the summer months is quite well related to the periods of fall ice phenomena in the estuary of the Amur River (Figure 52). As we see from studying the figure the relationship is in reverse, that is, the higher values of the sum of water temperatures correspond to the earlier periods of the appearance of ice phenomena. The reason for such a relationship between the periods of the ice phenomena and the sum of the water temperatures on the Korean cross-section is that, as was mentioned above, the higher ocean water temperatures should bring about an earlier change in the monsoon pattern from the summer to the winter type and a colder autumn and winter than for normal ocean water temperatures. The relationship given can be considerably improved if we use the air temperature at one or several continental points as a second variable characterizing the thermal condition of the continent. If we take the mean

temperature for September at Yakutsk the general coefficient of correlation representing the relationship of the periods of fall ice phenomena to the sum of the mean water temperatures for the summer months in the Korean Strait and the mean monthly temperature in September at Yakutsk, is exceptionally high:

$$R = 0.98 \pm 0.05$$

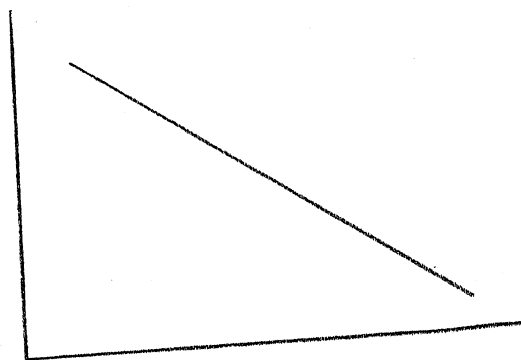


Figure 52. The relationship of the mean deviation from the normal of the periods of autumn ice phases at Cape Pronge and Baydukov Island to the heat content of the Tsushima current for May-August.

This relationship gives a 100 percent probability that the errors in forecasting will not exceed $1/5$ of the long range amplitude. Similar relationships were also obtained for the northern part of the Tatarsk Strait.

The shortcomings in the examined method are due, first of all, to the extremely limited number of observations of the water temperature in the Korean Strait. In increasing the number of observations some and maybe a considerable decrease may occur in the

coefficients of correlation and the reliability of the equations.

A general shortcoming in the examined method and those similar to it is that constructing a relationship of the type examined above it is necessary to strive for a possible big increase in the number of observations, since in the opposite case the relationships are not sufficiently dependable and require additional checking. Unfortunately, at the present time many prolonged observations either are not made for separate areas, or the observations made in different years are not uniform or of poor quality.

The necessity of using many observations arises because of the insufficient rigorousness of the physical basis of these forecasting methods. Actually, even though these methods are based on the correct physical prerequisites, a very large number of factors affecting the development of ice phenomena at sea are not computed by these methods. An evaluation of the importance of these factors is not made, and in many cases cannot be made. The proof of the small importance of the ignored factors is developed by purely statistical methods.

If in building a prognostic relationship 20 years were used and in this the unconsidered factors brought errors in the forecasts not exceeding the allowable limits, then it is supposed that even in the future the size and frequency of the errors will remain the same as for the duration of the studied period of time. From this the necessity of studying the largest number of observations possible becomes obvious, inasmuch as in just this case we can get sufficient assurance of the reliability of the prognostic relationship.

There is no doubt that all the relationships of the above types, widely used at the present time for long-range forecasts,

in the near future will be replaced by more complete forecasting methods, the basis of which will be computations founded on reliable long-range weather forecasts. But until then we cannot avoid using the long-range forecasting methods similar to those studied above, which currently are being used successfully in providing the national economy with forecasts of the ice conditions on the seas of the USSR.

FORECASTING THE LEVEL OF THE CASPIAN SEA FOR A YEAR IN ADVANCE

A good example of the possibility of building a method for long-term sea level forecasts without using weather forecasts is the method of forecasting the level of the Caspian Sea for a year in advance developed by G. P. Kalinin. This method is based on a careful study of the reasons causing a variation in the level of the Caspian Sea and the brilliant use of its individual peculiarities. The Caspian Sea is a tremendous closed-in body of water filled with salt water and is actually the largest lake in the world. The great importance of this sea to the national economy and the uniqueness of the patterns of this body of water were the reasons for the attention which was given and is currently being given by scholars to research work directed towards its study.

As far back as Peter I work was organized to study the Caspian Sea. The beginning of regular observations of the variations in the level of the Caspian Sea goes back to 1837, when a tide gage was established in Bakinskiy Bay. The first attempts, however, to study the variations in the level were made by Academician E. Lente in

1830. In 1884 a well known Russian geographer-climatologist A. I. Voyeykov made the first attempt to study the water balance of the Caspian Sea. Later much work on the study of the regime of the Caspian Sea was accomplished by N. M. Knipovich, A. I. Mikhalevskiy, S. Ya. Shcherbak, I. A. Benashvili, G. P. Ponomarenko, L. F. Rudovits, B. A. Apollov, G. R. Bregman. However, especially careful and fruitful studies in connection with the pattern of the Caspian Sea level were made in the State Hydrological Institute by a special group of scholars working under the direction of B. D. Zaykov. This study was undertaken to clarify the reasons for the exceptionally sharp drop in the level of the Caspian Sea that began after 1929, and in connection with the work of reconstruction on the Volga and other rivers flowing into the Caspian Sea. As a result of all these studies many extremely interesting peculiarities were established of which we will mention here only those which we need to know in order to understand the method of forecasting the variations in sea level.

The level of the Caspian Sea has a large annual and secular variation. In Figure 58 we have shown the changes in mean yearly sea level, and in Figure 53 we see the annual variation of the level. Of course, such large variations in the level have very great importance in ~~navigation~~ and for other human activities connected with the Caspian Sea. Drops in the sea level result in very extensive and expensive dredging of the canals used by ships as the approach to many ports located on the shores of this sea and at the mouths of major rivers, and also in the reconstruction of moorings in ports etc. When the level drops some fishing industries built on the shores of the sea are tens of kilometers from the water line since

considerable areas of the Caspian Sea are very shallow. A drop in the sea level caused losses of large areas in which many kinds of fish found food, etc. However, a rise in the level of the Caspian Sea is just as unfavorable to the economic activity of the different branches of the national economy as its drop. For the national economy the most desirable condition would be stabilization of the level. But, unfortunately, to assure stabilization of the Caspian Sea level complicated and costly hydrotechnical installations are needed.

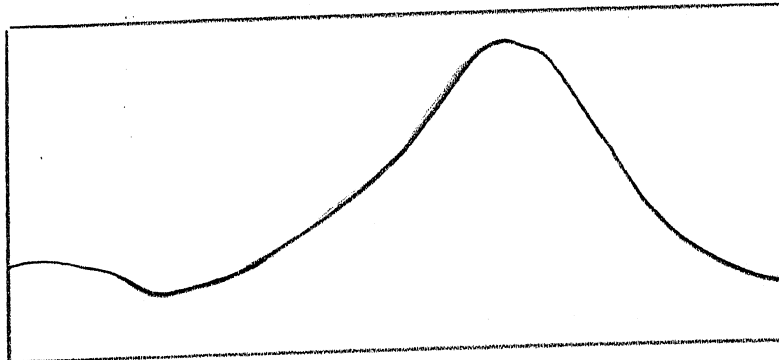


Figure 53. The annual movement of the Caspian Sea level.

What then is the reason for the sea level variations that bring so much annoyance to the activity of the economic organizations?

Several rivers empty into the Caspian Sea. These rivers bring tremendous amounts of water to the Caspian Sea which they collect in their basins from precipitations. The most voluminous river is the Volga. The discharge of the Volga River comprises 80 percent of all the water brought by all the rivers emptying into the Caspian. The greatest amount of precipitation falls into its

basin in the summer. But only 10 percent of the summer precipitations are carried to the sea. A great portion of these precipitations do not reach the sea but evaporate. In the winter there is considerably less precipitations than in the summer, but the winter precipitations are in the form of snow which evaporate hardly at all and do not seep through the ground, but are saved up for the spring air temperature rises. If the rise in temperature in the spring takes place very rapidly, the melting of the snow takes place very rapidly also. In this case only an insignificant percentage of the melted water has time to seep through the ground and to evaporate; a great portion of it runs into the sea in a comparatively short space of time. Conversely, in a slow, gradual rise in air temperature in the spring ~~and~~ the melted water seeps through the ground and evaporates more on the way to the sea, since the rise in river level is less than in the first case, and, consequently, the current speeds are less. All this causes large losses of melted water and the sea gets a smaller amount of water than in a rapid rise in air temperature in the spring, even when the amount of fallen snow in the river basin is the same. The fall air temperature regime also affects the amount of water flowing into the sea.

If the autumn was warm, the precipitations in the river basin were in the form of rain, and if it was cold, it was in the form of snow. In the first case these precipitations seeped off and evaporated, but in the second case they accumulated in the form of snow and evaporated slightly. Therefore, a cold autumn is attributable to an increase in the flow of water into the sea, and a warm one to a decrease.

The amount of water brought by a river to the sea is deter-

mined mainly by the winter precipitation. If by their temperature conditions the spring or the autumn is favorable to a large inflow then for a small amount of precipitation then no more water will seep off than the amount which fell in the basin. Also in case of a spring and fall temperature regime unfavorable for the inflow, the large amount of precipitation will only be decreased, but may still remain large despite the loss.

As can be easily established by the graph (Figure 53) of annual variations in the Caspian Sea level, the highest of sea levels are reached in July. From this we can conclude that the spring Volga flood succeeds in spreading out over the entire sea surface only in July. But the time, elapsing from the period of ice-melting, which occurs in the Volga basin in the middle of April, is used by the water on its course to the sea and on the spreading out over the sea.

In this manner the basic reason for the rise in the Caspian Sea level is clarified. The extent of the rise in the Caspian Sea level depends on the amount of water which comes from the rivers, chiefly from the Volga, during the spring flood. The lowest level during the course of a year in the Caspian Sea is in February and March when the influx of river waters is at a minimum. A decrease in the sea level in the period from August to January is due to the fact that the water flow from the rivers becomes a minimum during this period. However, evaporation from the surface of the Caspian Sea is observed throughout the entire year. A layer of a thickness of about 980 millimeters evaporates on a yearly average from the surface of the Caspian Sea is observed throughout the entire year. A layer of a thickness of about 980 millimeters evaporates on a yearly average from the surface of the Caspian Sea. Most of all

the water evaporates in the warm part of the year just at the period of time when the sea level reaches a maximum. However, at the same time the flow of melted waters into the sea reaches maximum values, ~~whereas~~ the amount of water entering the sea is greater than the amount of evaporated water. When, however, the flow of melted river waters into the sea subsides, and the waters derived from precipitation in the form of summer rain or from subterranean winter supply flow from the rivers, the influx of water is considerably less than the evaporation. As a result the level begins to decrease.

Studies of the water balance of the Caspian Sea showed that the evaporation from the surface of the Caspian Sea from year to year changes within comparatively small limits and, therefore, in the first approximation may be considered unchanged. We can also assume that the amount of precipitation falling directly on the mirror of the Sea to be unchanged from year to year. The latter assumption does not clearly correspond to actuality, but the amount of water falling on the surface of the Caspian in the form of precipitation is so small in comparison with other accountable forms of the water equilibrium (about 5 times less the amount of water coming from the river flow), that we may also disregard in the first approximation the variations in the amounts of precipitation falling on the mirror of the sea from year to year.

The next negligibly small element of the water balance is the inflow of subterranean waters directly into the Caspian Sea. It is evaluated only at 14 millimeter layer stretching over the entire sea surface. In this manner the river flow, which from the mean long-term data is computed to be an 82 centimeters layer lying on the

entire surface of the sea, is the basic variable value used in the water balance of the Caspian Sea. As was mentioned above the bigger part of the annual flow of waters -- about 63 percent -- enters the Caspian in the spring-summer period. That is why G. P. Kalinin attempted to determine the relationship between the spring-summer rise in sea level and the spring flow of the Volga River fixed along a line at the city of Stalingrad (Figure 54). This relationship was sufficiently close. Therefore, the author went further and established relationship between the precipitation in the basin of the Volga River for the period from September to March inclusive and the spring-summer rise in level. The change in the level from the lowest position of the level in February or March to the highest position of the level in July-August is called the spring-summer rise in level here. This relationship was somewhat worse than the relationship of the spring-summer rise in the level to the flow the Volga River, fixed at the city of Stalingrad. But this is natural. But then, when the flow on a closing line is taken, that amount of water which will enter the Caspian is computed by this with sufficient accuracy. It is no easy matter to compute the precipitation falling in the basin of as large a river as the Volga, instead of the flow. In order to shift from precipitation to flow, it is necessary to know the water loss in evaporation and seepage. The relationship shown in Figure 57 does not include the water loss; this explains the decline in the relationship which is seen in comparing it with Figure 54. Despite the decline this relationship still has a sufficiently high coefficient of correlation equal to 0.75. [See Figure 54 on following page].

Since the loss of water in the autumn and spring greatly de-

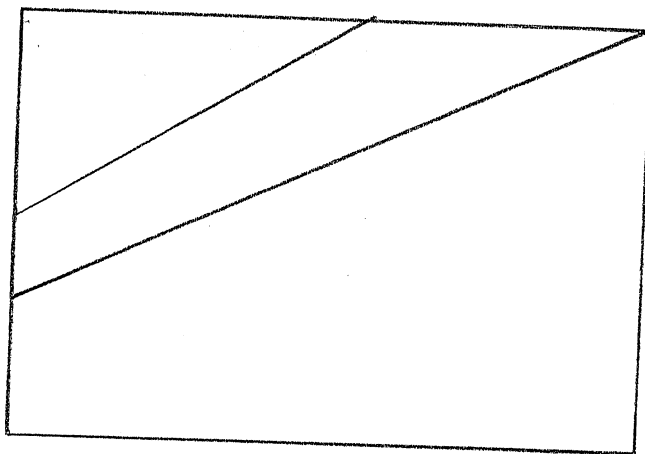


Figure 54. The relationship of the spring-summer rise in sea level to the spring (April-July) flow of the Volga River at the city of Stalingrad.

depends on the air temperature in these periods, G. P. Kalinin made an attempt to improve the relationship of the precipitation in the Volga basin to the spring-summer rise in sea level by inserting into this relationship the sums of mean monthly air temperatures for the autumn months of October-November and the spring month of March as a third variable. Inserting this variable he succeeded in increasing the coefficient of correlation from 0.75 to 0.85 = 0.05:

$$\Delta H = ay_{IX-XI, III} - bt_{IX-XI, III} - c$$

(a, b, c are the constant coefficients). The accuracy of the relationship is shown by the fact that 70 percent of all the forecasts from this relationship do not produce errors exceeding 5 centimeters, and the probable error is only ± 3 centimeters. In this manner it was possible to predict the maximum rise in the level, which is usually observed in July, from the information on the precipitation in the Volga basin for the period September-March, from the air temperature and from the position of the sea level in February-March. The forecast of the maximum level for the year from this relationship can be made in the beginning of April.

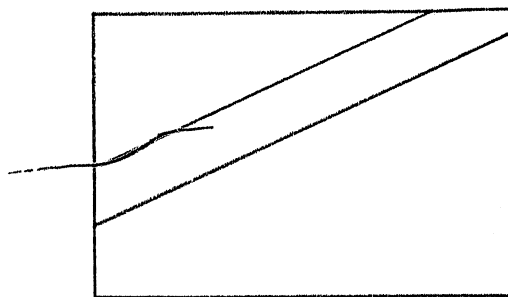


Figure 55. The relationship of the spring-summer rise in the level to precipitation (September-March) in the Volga River basin

Since the basic variable value from whose change the variations in the level of the Caspian take place, is the flow, and it is computed from the autumn-winter precipitations in the Volga basin when forecasting the summer maximum, we could assume the possibility of predicting the annual change in the sea level also from the summer maximum. This assumption is based on the fact that all the elements of the water balance of the Caspian Sea, except the flow, may be considered constant, with some degree of accuracy, while the main portion of the yearly flow of the rivers into the sea is determined by the spring flood, which is computed by the relationship examined above.

Bearing in mind what has been stated, G. P. Kalinin built a relationship between the annual change in the level and the height of the spring-summer rise in the level or, in other words, the summer maximum in sea level. The difference between the minimum sea level in two successive years, usually observed in February or March, is called the annual change in the sea level in the given case. This relationship is characterized by the coefficient of correlation $r = 0.89 \pm 0.02$ and is expressed by the following terms:

$$\pm \Delta h = A(\Delta H - b),$$

where A and b are the constant coefficients.

Using this relationship and knowing the forecast of the spring-summer rise in sea level, we can predict the minimum sea level, which will be observed in the following year in March. In this manner the combination of the two relationships allows us to predict the sea level a year ahead.

Inasmuch as the sea level from the minimum to the summer maximum and from the summer maximum to the minimum of the following year changes smoothly, we can, without any difficulty, interpolate the level for each month of the year basing this on knowledge of the actual March sea level in the current year, of the predicted maximum sea level in July and of the predicted minimum level for March of the following year. The forecasts of the Caspian Sea level proposed by G. P. Kalinin are given beginning with 1938 up to the present time. In Figure 56 we have shown the course of the actual and the predicted mean annual sea levels. As we see from this graph the level forecasts for all these years were justified rather well. The level forecasts for the entire year are usually made in the beginning of April of each year. Thereupon, when the level reaches a maximum (this is usually observed in July-August), by the relationship of the annual change in sea level to the spring-summer rise we get a more precise level forecast for the remaining period of time to March of the following year inclusive.

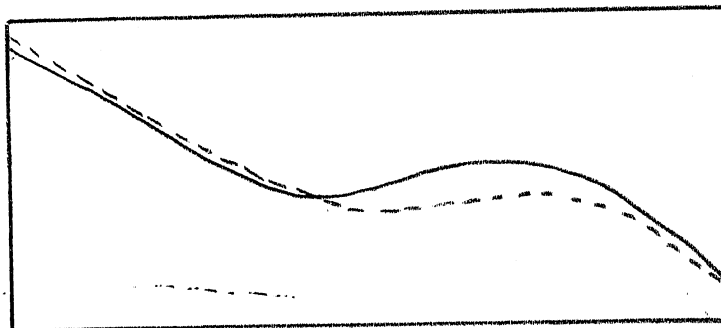


Figure 56. The course of the actual (1) and the predicted (2) mean annual values of the Caspian Sea level (Cape Bailov at the city of Baku)

The described method, proven in operational work, allows us to service the interested organizations with forecasts of the Caspian Sea level a year in advance. When the method was first applied many doubted the possibility of such a long-range forecast of the sea level. But ^athorough familiarity with this method convinces us of its adequate reliability. There is no doubt that this method could be improved considerably if we could use the long ranged forecasts of the precipitation for the summer period. The errors in the course of the level predicted by this method are often connected with the fact that the precipitation for the summer period is not computed at all. However the coefficient of the summer flow is small; but with large amounts of precipitation falling into the Volga basin in the summer the actual level of the Caspian Sea is, as a rule, greater than the one computed by the examined method and, conversely, in cases of exceptionally small amounts of precipitation in the Volga basin the computed levels are higher than the actual ones. With trustworthy long-range weather forecasts, probably it will be best to compute also the changeability of the evaporation and the changeability of the precipitations on the sea mirror. Having made the method more precise one can probably count on considerably smaller errors, and even failing that on not too large ones, in the forecasts obtained by applying the method described here.

39. THE METHOD OF FORECASTING THE VARIATIONS IN THE LEVEL OF THE CASPIAN SEA FOR 5-6 YEARS AHEAD

To bring into practice the servicing of the national economy with the above studied method of forecasting the Caspian Sea level for one year ahead did not nearly satisfy all the demands of the economic organizations. Requests for forecasts of the Cas-

pian Sea level for several years ahead were and are being made to the Central Institute of Forecasting of the Hydrometeorological Service. These demands in most cases are in connection with the need of projecting and constructing major installations on the shores of the Caspian Sea. Unfortunately, until the present time most of these demands could not be satisfied, even though in the past few years one other method of extra long-range forecasts of the Caspian Sea level was brought into operational practice in the Central Institute of Forecasting permitting prediction of the mean annual levels for 5-6 years ahead. Nevertheless, the introduction into practice of these forecasting methods did not manage to satisfy fully the requirements of economic organizations, since in order to project major installations it is necessary to know which levels of the Caspian should be expected in 20, 30 or 50 years.

There is no doubt, that even forecasts for 5-6 years are necessary to the national economy, they will play a positive part in the matter of warning about possible level changes in the next few years. As an example of the use of such a forecast we can use this fact. If a forecast for 5 years anticipates a possible considerable increase in the level for this period of time, then the economic organization planning the construction of ships for dredging in the northern Caspian may take into account the forecast of the level and, depending on it, build a greater or smaller number of ships for this purpose. We can give many similar examples. To illustrate the economic significance of these forecasts we may point out that for a level drop of 5 centimeters about one million rubles are spent merely on dredging the Volga-Kaspiysk Canal.

Attempts to predict the Caspian Sea level for prolonged periods of time were undertaken a long time ago. Even Ye. Bryukner supposed a high probability of the 35-year periodicity in the changes in the level of the Caspian G. P. Bregman and other researchers proved that in the variations in the level of this sea a 35-year periodicity is lacking, but that the variations in the level have a cyclic nature. It is apparent that an increase in the report period of the forecast in comparison with that which is given by the method of G. P. Kalinin could have been obtained only by studying large-scale atmospheric processes. Unquestionably, an increase in the report period of the forecast cannot be attained by taking into account the preceding condition or the course of hydrometeorological elements in the basin of the Caspian Sea. First of all it was necessary to determine the reasons for the exceptionally sharp drop in the sea level after 1929 when in 15 years the sea level dropped almost 2 meters.

L. S. Berg, studying the historical data on the variations in the Caspian Sea level parallel with the data on navigation in the arctic seas, came to the conclusion that a decrease in the level of the Caspian occurs approximately in the same years when light ice conditions are observed in the arctic seas. As we know, many researchers pointed out the exceptional heating of the Arctic observed after 1920. Other researchers, however, working in this field, considered that the heating of the Arctic was clearly noticeable only after 1932. Whichever way it is, even in our time the heating of the Arctic can be considered coincident with the sharp drop in the Caspian Sea level. As one of the first, V. Yu. Vize pointed out that the reason for the heating of the Arctic is the intensification of

the general circulation of the atmosphere. Later this idea of V. Yu. Vize was shared by other scholars too.

There is no doubt that the drop in the Caspian Sea level is a process very clearly characterizing the climatic changes occurring on a large territory. Studies by P. S. Kuzin of the variation in air temperature and the precipitation for the period of the sharp drop in the sea level underwent a considerable change. The heating of the Arctic also is not a local phenomenon. It can occur only as a result of very considerable changes in the course of atmospheric processes over a large territory including if not the entire world then in any case the northern hemisphere. All this taken together forced the authors of the method of forecasting the variations in the Caspian Sea level examined here -- N. A. Belinskiy and G. P. Kalinin -- to study the peculiarities of the atmospheric processes causing such remarkable changes in the level of the Caspian Sea.

For evaluating the development of the atmospheric processes they used the index of atmospheric circulation introduced by Belinskiy. An analysis of the atmospheric processes in connection with the variations in the Caspian Sea level could be made only from the products of observations for a prolonged period of time. In order to avoid time consuming and tedious work they used the data available in the synoptic catalogue. To simplify the calculations they adopted a scale in balls for the evaluation of atmospheric processes shown below:

-5 is a powerful anticyclone (the middle isobar is 1035 millibars and more);

-4 is an anticyclone of medium intensity (the middle isobar is 1025-1030 millibars);

-3 is a weak anticyclone (the middle isobar is 1020 millibars or less);

+5 is a deep cyclone (the middle isobar is 990 millibars or less);

+4 is a cyclone of medium intensity (the middle isobar is 995-1000 millibars);

+3 is a weak cyclone (the middle isobar is 1005 millibars and up).

By this scale a daily evaluation was made of the atmospheric processes for 8 areas into which were divided the western part of the Atlantic, Europe and Western Siberia for the period from 1900 to 1945. To characterize of any period of time they used algebraic sums of the balls computed for each area separately. Belinskiy undertook the study of the changes in the index (the sum of the balls) and its comparison with the course of hydrometeorological elements. In this it became clear that the index of atmospheric circulation in general reflects the characteristic signs of the variations in the Caspian Sea level and also agrees with the changes in the iciness of the Barents Sea. Besides this, the specular course of the changes in cyclonic and anticyclonic activity over the ocean and land and over the northern and southern areas of one hemisphere in the annual as well as the long-range cross-sections was noted. Further studies made by N. A. Belinskiy simultaneously with G. P.

Kalinin showed that the amount of precipitation falling in the Volga basin is related to the intensity of cyclonic and anticyclonic activity over the continent. A greater cyclonic activity over the continent accounts for a greater amount of precipitation, and a greater anticyclonic activity for a smaller amount of pre-

precipitation. The increase in the atmospheric circulation observed for the past few years was expressed in the remarkable intensity of cyclonic activity over the ocean and arctic seas. For the same period of time anticyclonic activity over the continent grew considerably. A natural result of this process was the decrease of precipitation in the Volga basin, and, consequently, in the basin of the Caspian Sea too. Together with this, as we know, an increase in the cyclonic activity in the northern areas and over the arctic seas brought about an increase in the air temperature in these areas, and as a result of the increase in the air temperature the iciness decreased. In consequence of the reduced precipitation in the Volga basin a decrease in the Caspian Sea level occurred. While following the increase in air temperature in the areas of the arctic seas the iciness decreases. In this manner they succeeded in explaining the change in the iciness of the arctic seas mentioned by L. S. Berg and the parallel change in the Caspian Sea level. The reason was the increase in the atmospheric circulation which manifests itself by an increase in the cyclonic activity over the ocean and by an increase in anticyclonic activity over the continent.

Besides this, in the cyclonic and anticyclonic activity cyclic variations were noted. These variations in the annual and long-range cross-section pass over the continent and the ocean with phases differing by 180 degrees. This means that during the time when a cyclonic activity develops over an ~~ocean~~, an anticyclone exists at the same time over the continent. Conversely, if anticyclones are developing over ^{an} ocean, then a cyclogenesis exists at the same time over the continent. Cyclonic variations in atmospheric processes are characterized by the fact that it is distinguished from

other cycles by the amplitude as well as by the period of the cycle, having a mean period of about 12 years. If the full cycle of the variations is about 12 years, then half of its period -- 6 years -- will be corresponding to 180 degrees, that is, exactly that value by which the variations in the intensity of the cyclonic and anticyclonic activity over the ocean are ahead of similar variations over the continent.

On the basis of these considerations a graph of the relationship between the sliding 5-year sums of the indexes representing the intensity of the cyclonic and anticyclonic activity in the area of the Azor maximum and mean flow of the Volga River for 5 successive years was constructed. In other words, if the index was taken for 1901-1905, 1902-1906 etc. then the flow of the Volga was averaged for the years 1906-1910, 1907-1911 etc. and the corresponding points were plotted on the graph. (Figure 57). This figure shows that the relationship between the intensity of the cyclonic and anticyclonic activity in the area of the Azor maximum and the flow of the Volga is^a comparatively close one. (the coefficient of correlation $r = -0.84 \pm 0.05$), whereas the circulation index is 5 years ahead of the Volga flow. Consequently, using this relationship, we can predict the Volga flow 5 years in advance. But inasmuch as the Volga flow is related to the atmospheric circulation we can consider that the Caspian Sea level is also closely related to the atmospheric circulation, since it is the Volga flow which determines basically the extent of the variations in the Caspian Sea level. Comparing the sea level variations with the changes in the circulation indexes we must bear in mind the nature of the relationship of the level to the flow. In the Caspian Sea we get an addition of the

river flow from year to year, but to deduce the expenditure of water in evaporation:

$$H_t = H_0 + \sum_{t=0}^t (P - R) \Delta t.$$

Here H_t is the level at a moment of time t ; H_0 is the level at a point in time t_0 ; P is the influx of water for a unit of surface of the sea mirror (precipitation, flow); R is the expenditure of water (evaporation, filtration) from a unit of sea surface.

millimeters

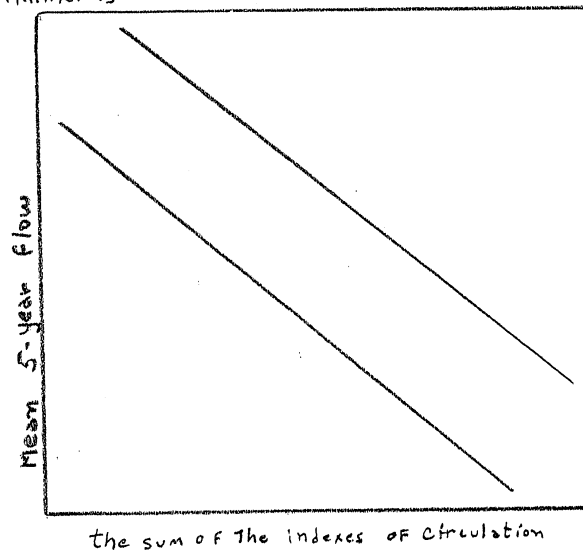


Figure 57. The relationship of the mean 5-year values of the flow at the city of Stalingrad to the sums of the correlation indexes for 5 preceding (in relation to the flow) years (the sum of the indexes is decreased 100 times).

In this manner insignificant variations in the flow observed in nature from year to year are ironed out in the process

of forming the sea level. Large cyclic processes (with a mean period of around 12 years) are not evened out by summation, but remain just as relief-like as before summation, though more clear as a result of the evening out of the small variations. The deviation of the flow from the norm, keeping the sign for a considerable time, cause either a decrease in the level or an increase. Therefore, it is natural to assume that the condition of the level for which there is no decrease or increase in sea level depends on some value of the index of atmospheric circulation, that is, the influx part of the water equilibrium of the Caspian Sea is equalized by the outflow part. The deviations, however, of the index of circulation from the norm with one sign must correspond to the increase or decrease in sea level. If all this is true and the relationship between the index of circulation and the flow of the Volga is linear, then the resulting summation of the deviations of the index of circulation from the norm (corresponding to a constant level) should give values corresponding to the variations of the Caspian Sea level.

In view of what has been said the integral values of the deviations from the norm of the index of atmospheric circulation were computed and a relationship of these values of the index to the Caspian Sea level was set up. The relationship of the integral values of the circulation index in the area of the Azov maximum to the level of the Caspian is characterized by the coefficient of correlation $r = 0.95 \pm 0.02$ when the circulation precedes the level of the Caspian by 5 years, but when it moved to 6 years relative to the sea level the coefficient of correlation was still higher $r = 0.96 \pm 0.01$. The equation expressing the relationship of the Caspian Sea level to the circulation has the following form:

$$H_n = 352 - 0.083 \sum_{1900}^{n-6} (i_2 + 320).$$

Here the following conventional designations are made: H_n is the Caspian Sea level of the corresponding year: n is the number of the year for which the level is being computed; i is the index of circulation.

Using this equation the mean annual values of the sea level for each year were computed from the corresponding values of the index of circulation and compared with the actual movement of the Caspian Sea level for a long period of time (Figure 58). As we see from the figure the variation of actual and computed level values coincides rather well: the maximum error does not exceed 32 centimeters, and the average error is only 16 centimeters.

In 1946 with the aid of this relationship the first forecast of the level change for 5 years was made and in the future this forecast will be made yearly. The actual values of the Caspian Sea level predicted with a 5-year report period was sufficiently close. [See Figure 58 on following page]

The above examined examples of the methods for ice and level forecasts variation show that one must make an analysis of hydro-meteorological processes in order to obtain relationships necessary for making long-range forecasts of some elements of the hydrometeorological regime. These examples show that despite the physical basis of the construction of long-range hydrometeorological forecasting methods it is in many cases inadequate. In connection with this there arises the necessity of verifying by statistical methods, using a large number of observations, the logical conclusions based on physical laws.

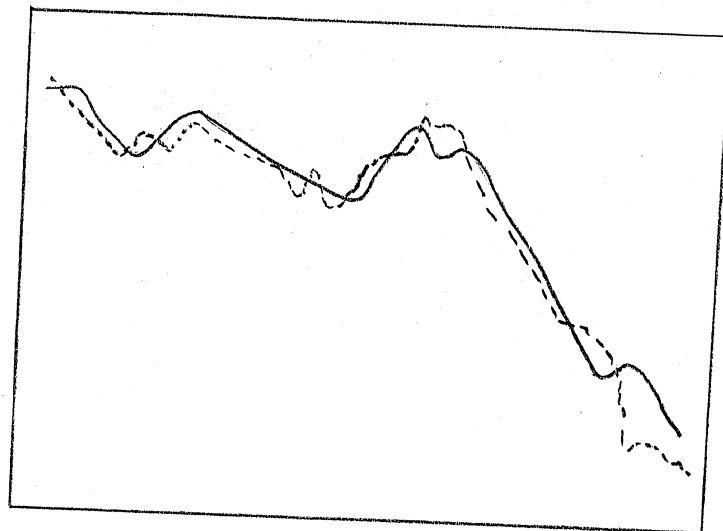


Figure 58. The long-range movement of the actual (1) and computed (2) mean annual values of the Caspian Sea level (at Baku):

$$H_n = 35.2 - 0.083 \sum_{1900}^{n-6} (i_2 + 320)$$

CHAPTER VIII

THE REQUIREMENTS FOR ESTABLISHING AND CONDUCTING MAR-
INE HYDROMETEOROLOGICAL OBSERVATIONS

For the proper and effective servicing of the economic organs and the defense establishment of our country with information and forecasts of sea condition we must, first of all, set up a proper and well-organized system of observations. Most of the forecasting methods presented above are based on a careful analysis of observation products. If the observations used for working out forecasting methods for some reason were insufficient, then the forecasting methods when applied in practice cannot give positive results. That is why the quality of observation must be completely irreproachable.

The quality of the observations is related to a large number of factors. An apparatus with the aid of which the observations are made may be entirely unsatisfactory. It may be that the apparatus is perfectly adjusted but was calibrated long ago (for example, an air vane or anemometer) -- the products of observations from such an apparatus in the end will also prove at variance with actuality. In many cases the quality of the observations depends on the ability of the observer to complete a job assigned to him, on the care with which the observations are made. All the factors on which the quality of the observations depends are adequately and fully presented in appropriate instructions and directions. Therefore, there is no need here of dwelling especially on them. We should note the main questions of principle in establishing and conducting observations which have become especially acute in connection with the development of forecasting methods. The hydrometeorological observations

made a long time before the beginning of the scientific development of the marine forecasting methods in many cases are either insufficient for the development of the methods or are entirely unsuitable for these purposes. The reason for this is that in setting up the observations there was no clear idea of how and for what purpose they were to be used later. The object of the observations was to give the most general picture of the hydrometeorological regime of the sea or its separate areas. In short, the object of the observations was to elucidate what is on the sea. By accumulating the observations the great changeability of the pattern of different elements was clarified. In connection with this, in order to characterize the regime, one began to compute the mean and extreme values, frequencies and other statistical characteristics observed in the movements of different elements of the regime. Up to the present time much work has been done on characterizing the hydrometeorological regime of the seas and oceans by statistical methods. This work is summed up in a number of articles and textbooks. Now most of the seas are classified with some degree of reliability by statistical information on the hydrometeorological regime. At present there still remain some blank spots either entirely unclarified by information on the regime or characterized by few observations.

Simultaneously with the accumulation of observation products, the physical nature of the phenomena observed at sea was studied by physiomathematical analysis. The physics of the sea was created.

Aside from statistical characterization of the seas regime a qualitative physiogeographical analysis was also made, and oceanography was created. Those phenomena which were characterized by

data obtained from observations were explained by regularities established by physiomathematical analysis.

At the present time knowledge of the oceans and seas has expanded considerably, and products of observations have been accumulated. The problem of what is observed on the ocean and seas, what phenomena may be encountered in any area, generally speaking may be considered solved for the greatest majority of the areas of the world's oceans. Now the research navigators have a more complex problem -- studying the phenomena and processes occurring in the ocean and seas in their interrelationship and interdependence in the concrete geographical situation in which they are observed. In other words this problem may be formulated as the enunciation of general laws of marine physics for concrete geographical conditions. The problem of developing the methods of marine hydrometeorological forecasting is solved simultaneously with the solution of this problem.

A new problem requires new methods for its solution, and consequently, new methods of observations.

If in the past every visit to any ocean or sea area yielded observation products of exceptional value, then the production of particular observations in many areas of the ocean and seas in most cases only serves to verify the statistical characterization of the regime. There is no doubt of the necessity for further perfecting and defining the statistical characteristics of the regime of ocean and seas, but this problem is not the main one now and can be successfully solved by means of shipboard and aircraft routine observations. Special hydrometeorological stations, observatories and expeditions

must radically reorient their work toward establishing and conducting observations. They must set and conduct their observations so that in the shortest possible time they could solve the problem: what are the interrelationships and interdependences of the elements of the regime in the concrete physiogeographical situation of any area of the ocean or sea. The observations should be set so that they can best provide navigation with information on the current sea condition (in a wide sense of the word) and with forecasts of the sea condition in the future.

It is necessary to explain this by examples.

Above we examined problems connected with forecasting currents. In this it became clear that an orderly theory on drift currents cannot, by far, always be used in pure form for computing the currents in a certain marine area for a given wind. It helps to break down complicated phenomena observed in nature. Still, the coefficients relating the current to the wind causing it must be determined from observations.

How were observations made before?

If we disregard the floating lighthouses from which observations were made regularly we can discuss only the single observations in seas without tides and the semidiurnal and daily observations in seas with tides, as a rule made in calm weather. It is quite obvious that single observations of currents cannot be used in building relationships and in determining the necessary wind coefficients. These observations are suitable only for the statistical characterization of currents, and then only for weak winds.

How are we to make observations of currents?

Two ways are possible: either the observations must be continuous and over a long period of time, or the observations must be made applicable to the definite and specific characteristic peculiarities of the wind regime, that is, they must be made in large quantity and for various wind directions and speeds, including even the high wind speeds. The first way has great advantages over the second one, they are as follows: (1) In conducting lengthy continuous observations peculiarities of the regime will be brought out which may be omitted by individual observations; (2) the current may be studied for a shorter period; (3) in completing them at several points it is possible to compare the pattern of currents observed in various areas; (4) only by this method of observation do we get sound information on the actual condition of the current at sea, and also a check on the forecasts of currents without which it is senseless or simply impossible to compile forecasts of currents.

How may we make observations of this type?

In our time when the science of automatic machines and telemechanics underwent unprecedented development, we should make wide use of self-recorders of currents and automatic instruments, transmitting by radio, telephone or telegraph the information gathered by the instruments. However, the installation of such equipment is possible only in a coastal area and at comparatively small depths. In deep water areas, however, these observations must be made from ships operating for a long time in definite areas or systematically cruising along an established route.

The same is true with the observations of other elements of

the regime. Unsystematic observations of temperature and salinity do not permit us to make a presentation of the changes in the thermal and salinity pattern of the sea even during its actual existence over a certain period of time on a considerable area of the sea.

As was already said in the section on forecasting ice phenomena in an open sea, it is necessary to make systematic observations established once for many years. These observations must be made annually on the same dates. The frequency with which these observations are made cannot be standard for all seas and all areas. It must be determined depending on the steadiness in time of the hydrological pattern in each area. The steadier in time the hydrological pattern is, the less the frequency of making the observations on hydrological cross-section. Unfortunately, at the present time there is still much time and effort being spent on unsystematic hydrological work, but standard observations allowing us to watch the changes with time in the pattern are made very rarely. Also, no work is being done which would allow us to establish firmly the frequency of the observations on standard cross-sections. The frequency of observations is determined by departing from formal concepts. For example, the periods for making observations on cross-sections are established as once a month or once a season; it remains uncertain why just these periods are selected. To understand this better it is useful to refer to an example again. When we speak of the air temperature the measure of the temperature for a month is taken as the mean monthly temperature and from this value we judge whether the month was cold or hot. But the mean monthly air temperature is computed as the mean of the mean daily temperatures, and the latter is computed from four observations made at various hours of the day. All these computations or even several of them as a measure of the

air temperature for the month. To what degree we may characterize the temperature and salinity of water by one observation a month in an open sea without special studies of this problem for different physiogeographical conditions would be impossible to decide.

A great weakness of ice observations is that in observing the ice such elements as the amount of ice, the density of ice, the width of the land floe, the ice drift, are determined by sight. It is difficult to use these observations in building the corresponding quantitative relationships, as they are inexact and subjective. Weaknesses of visual observations are aggravated by the frequent change in instructions for making these observations. It is necessary to change over to instrumental observations in the shortest possible time. There are no difficulties in using rangefinders and angular rangefinders in determining the width of a land floe, the ice drift along a shore etc. In determining the amount and density of ice, however, we can successfully use optical instruments especially designed for this purpose. On the introduction of instrumental ice observations largely depend on the quality of the ice forecasts which have extremely great importance in navigation. Inasmuch as instrumental observations are at the present time fully available it is necessary to adopt them as quickly as possible.

The observations over the ice during ice air reconnaissance, having received a great lift in the past few years and in many ways aiding the better study of the ice regime of seas, unfortunately, are also made visually, even though here there are great possibilities of making instrumental observations by means of optical instruments.

The observation of disturbances is in an exceptionally poor

situation. Until the most recent times in many cases observations at shore and shipboard stations were not made of wave elements but an evaluation of the disturbance was made in balls. It is necessary to break away completely from the evaluation of a disturbance in balls and to change over completely to observing the wave elements even when these observations must be made visually.

And, finally, the last problem on which we must dwell is that of the representativeness of the observations made by shore stations for the area in which the stations are located. Actually, if we recall the relationship, examined above, used in forecasting water temperature or the periods of the appearance of ice, it will become clear just how essential a part the representativeness of the sea surface temperature that is observed by a station plays for the water surface temperature of the entire area for which the forecast is made. If the temperature of the water surface is observed at a steep shoreline, these observations in most cases will represent the temperature of the water surface observed on a large sea area adjacent to this area with similar depths. If, however, the observations are made at a shallow shore, these observations will represent the temperature of the water surface only in shallow areas. In deep water areas in the warm part of the year the water temperature, as a rule, will be less, and in the cold part of the year it will be higher than that observed in shallow areas. Consequently, in forecasting the periods of the appearance of ice from the temperature observed over a shallow area, the appearance of the ice can be correctly computed only in shallow areas. For deep water areas, however, the computation of the period of the appearance of ice may be unsuitable in practice. This applies also

to the forecasts of variations in water temperature.

In connection with this it is necessary for each station, when determining the place for making water temperature observations, to take into consideration the significance of these observations for the given sea area; of a bay, harbor, etc. If the interested organizations must have forecasts of the water temperature in a shallow water area. If forecasts of the appearance time of ice on the approaches to a port are required, then the observations of the water temperature are best made at a deep offshore area or from the head of a breakwater, etc.

In this way we should deal thoughtfully with the setting up and conducting of observations, complying both with the peculiarities of observations which must be considered in developing a forecasting method and with the demands of the organizations interested in the forecasts.

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Noted errors: [Errors pointed out on sheet attached to original manuscript have been corrected. Corrections were made and incorporated with this typewritten manuscript in their proper places]